THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR .... Jan Michael Bednarski ....

TITLE OF THESIS .... Holocene Glacial and Periglacial ....

Environments in the Whistlers Creek ....

Valley, Jasper National Park ....

DEGREE FOR WHICH THESIS WAS PRESENTED .... Master of Science ....

YEAR THIS DEGREE GRANTED .... 1979 ....

The Upper Whistlers Creek Valley

THE UNIVERSITY OF ALBERTA

HOLOCENE GLACIAL AND PERIGLACIAL ENVIRONMENTS IN THE WHISTLERS CREEK VALLEY, JASPER NATIONAL PARK

by

© JAN M. BEDNARSKI

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE

OF MASTER OF SCIENCE

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA

SPRING, 1979

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Holocene Glacial and Periglacial Environments in the Whistlers Creek Valley, Jasper National Park.

submitted by Jan M. Bednarski

in partial fulfilment of the requirements for the degree of Master of Science

This thesis is dedicated to my Mother and Father.

ABSTRACT

The development of Holocene glacial and periglacial features in the upper Whistlers Creek Valley, Jasper National Park, was investigated through the use of airphoto analysis, detailed mapping and field research. Relative age dating techniques such as: lichenometry, dendrochronology, surface weathering, and the basic stratigraphic techniques of cross-cutting and superposition were employed.

The head of Whistlers Creek Valley is occupied by two cirque glaciers which show evidence of recent retreat. Both glaciers have active tongue-shaped rock glaciers immediately downvalley. Within one of the rock glaciers large thaw pits and exposures of foliated ice, dipping upvalley, suggest buried glacier ice. Debris supplied by avalanches and rockfalls are considered the major causes of burial. Tilted trees indicate that one of the rock glaciers has advanced at least 1 m in the last 45 years. Superimposed lobes on both rock glaciers suggest two periods of recent activity. These periods have been lichenometrically dated using *Rhizocarpon geographicum* at ~1770 and 1885 AD for one rock glacier, and older than 1700 AD and 1925 AD for the other. Evidence for a pre-Little Ice Age glacial advance is provided by a large lateral moraine in one of the cirques (lichenometrically dated >850 BP).

An inactive spatulate and several lobate rock glaciers were also identified. Total lichen cover, and rock weathering criteria, have established that these relic rock glaciers formed during a substantially older period. Dateable organic and tephra deposits, however, were not

observed. As opposed to the rock glaciers in the cirques, these older rock glaciers have developed beneath scree slopes and interstitial ice may have been responsible for their movement. A thin discontinuous drift sheet and erratics provide evidence of a Cordilleran Ice advance in this valley from the west.

Lichenometry was particularly useful for the minimum dating of the more recent (Little Ice Age) glacier and rock glacier activity. However, on surfaces older than 100 years, isolated lichen thalli become exceedingly hard to distinguish making measurement impractical. Problems of species succession and senescence also become evident on the older surfaces. Oxidation rinds were the only rock weathering criterion which could be systematically measured, owing to the resistant nature of the quartzitic rocks in the study area. Three soil pits located on surfaces of differing age, indicated slight variations in horizon development and clay mineralogy.

Turfbanked and stonebanked lobes and terraces, polygons, frost boils and sorted stripes are widespread in the Whistlers Creek Valley. A number of solifluction forms show evidence of reactivation. Gliding blocks up to 4 m in length are ploughing gentle, vegetated slopes leaving furrows up to 20 m in length. The size and significance of these blocks warrant further research.

ACKNOWLEDGEMENTS

Many people have helped me throughout the making of this thesis. Firstly, I would like to thank John England, without whom I would not have started or finished this thing in the first place. I would also like to thank my supervisory committee for their constructive criticisms and the many people whom I have consulted at one time or another. Raynor Shaw's help in surveying the rock glaciers was much appreciated. I would also like to give special thanks to Mike Ostafichuk for identifying the lichens, Graham Spiers for advice and hours of work processing the clay minerals, and Anita Moore for typing the manuscript.

Financial assistance from the Boreal Institute for Northern Studies is gratefully acknowledged. Lastly, I would like to thank Parks Canada for allowing me to work in Jasper National Park, and my Morris for getting me there.

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CHAPTER ONE

INTRODUCTION

1.1 Nature of the Study

In the Rocky Mountains of Alberta, climatic variations during the Holocene (10,000 BP to the present) and their environmental effects are poorly documented. Studies on this subject have been principally limited to reconnaissance surveys and inventories of prominent glacial features occupying major valleys. Detailed work, which until recently has been confined to these valleys, is usually centered on the longer time intervals of the late Pleistocene.

The purpose of this study was to investigate the climatically induced geomorphic history of Whistlers Creek Valley during the Holocene. This was done by investigating landforms in the valley which were produced by fluctuations in glacial and periglacial systems. A chronology of these fluctuations was determined using a number of relative age dating techniques. However, as with most studies of this kind, an incomplete understanding of how glacial and periglacial systems react to climate limits paleoclimatic reconstructions. This complication, and lack of absolute dates for geomorphic events, present problems which may not be solved. Nevertheless, the landforms in the study area are described and a general scheme of Holocene geomorphic events is presented.

1.2 Previous Studies

In this section some of the paleoclimatic studies done in the

Banff and Jasper National Parks area are reviewed, as well as some of the major works for the Cordilleran zone of the western United States. A number of techniques used are noted and chronologies of glacial events are presented.

1.2.1 Banff-Jasper area

The work done in the Banff-Jasper area (Fig. 1.1) may be divided into late Pleistocene and Holocene studies. As previously mentioned, the majority of studies done on climatic fluctuations have been concerned with the late Pleistocene geology of main valleys.

A. Late Pleistocene Studies

Discussion of late Pleistocene glaciations is primarily concerned with the deglaciation of the Banff-Jasper area as well as the large scale modifications of alpine topography by the ice. Rutter (1966, 1972) recognized four major glacial advances in the Bow River Valley, Banff National Park, during the late Pleistocene. His interpretation is based on tills and glaciofluvial deposits together with geomorphological and crosscutting relationships. North of the Bow River, Harris and Boydell (1972) found only a single till from a major glaciation in the Clearwater River Valley, and deposits from three glacial advances in the south branch of the Ram River Valley. Further north, McPherson (1970) described two late Pleistocene advances in the upper North Saskatchewan River. The present landscape is the result of the last advance, the Main Advance. During the retreat of this glaciation, two stillstands,

FIGURE 1.1 The Jasper - Banff area.

or minor readvances, are inferred from end moraines at Saskatchewan Crossing and at the junction of the North Saskatchewan and Alexandra Rivers. Westgate and Dreimanis (1967) have radiocarbon dated a charcoal layer near the base of a loess deposit which overlies ice-contact glacio-fluvial sands and gravels at a site near Saskatchewan Crossing. This date of 9330±170 BP provides a minimum estimate of late Pleistocene deglaciation for the upper North Saskatchewan River Valley. Rutter (1972) feels that the last glacial advance in the North Saskatchewan Valley is correlative with his last, the Eiesenhower Junction, advance in the Bow Valley. Both of these advances, therefore, are probably older than 9330±170 BP. However, Rutter (1976) also correlated his Eiesenhower Junction Advance with his Deserter's Canyon Advance in the Peace River Valley in northeastern British Columbia which has been dated between 9280±200 and 7470±140 BP. This apparent contradiction of dates may be explained if the charcoal layer at Saskatchewan Crossing has been reworked, as sometimes occurs (Rutter, Pers. Comm., 1977). On the other hand, recent work has found evidence of only limited ice advances in the mountains at this time. Osborn and Duford (1976) have identified a post Wisconsin, but pre 6600 BP, glacier and rock glacier advance based on volcanic ash evidence. Luckman (1978) obtained basal dates, from peat deposits, of 6920±100, 7770±110, and 8110±100 BP from the Sunwapta Pass, near the headwaters of the North Saskatchewan River and a date of 9660±280 BP from the Tonquin Pass, near the continental divide. These dates suggest that the upper mountain areas were ice free by ~10,000 BP.

Roed (1975) studied the contact zone between Laurentide and Cordilleran tills in the Hinton-Edson area. He found evidence of two major Cordilleran advances which flowed well out of the mountains. Pauses in the recession of the last major advance, the Obed, are associated with kame deposits and the development of two river terrace levels in the Athabasca River Valley. After the major advances there was also a minor advance, the Drystone Creek Advance, which was confined to small valleys in the Front Ranges and foothills of the Rocky Mountains. Roed (1975) feels that the Drystone Creek Advance is correlative with Rutter's (1972, 1976) Eisenhower Junction and Deserter's Canyon Advances. He also assumed that his entire Cordilleran piedmont sequence falls within Richmond's (1965) Pinedale Glaciation of the western United States.

B. Holocene Studies

Before discussing the Holocene studies, the terminology for the Geologic-climate intervals of the Holocene is reviewed. The interval of "mild climate" (optimum?) which followed the last major glaciation will be referred to as the "Hypsithermal" as defined by Porter and Denton (1967, p. 180). The subsequent interval, characterized by cooler climate and glacier expansion after maximum recession during the Hypsithermal is referred to as "Neoglacial" (Porter and Denton, 1967, p. 181). This interval generally began 5000-3000 BP. The term "Little Ice Age" was initially defined by Matthes (1939, p. 520) and represented the same time span as Porter and Denton's "Neoglacial". Common usage, however, has since restricted the meaning

of this term to denote the general cooling in climate from the early 16th through to the 20th Century. This period is characterized by the rebirth and expansion of alpine glaciers (Denton and Karlen, 1973, p. 156) and will be used here in its restricted sense.

Heusser (1956) wrote an extensive paper dealing with postglacial (Holocene) environments of the Robson, Jasper, Banff and Yoho National Park areas. His scheme of environmental fluctuations was primarily based on dendrochronology and dealt with the Little Ice Age. By coring ice-thrust trees he was able to date the positions of the ice front. He studied a total of twelve glaciers, and, in summary of the glacial data, Heusser noted that all the variations between the twelve glaciers appeared synchronous. Substantial advances occurred in the last few centuries with maximum positions being attained between the late 17th and early 18th Centuries. The ice fronts remained stable until the late 18th Century, after which, recession occurred and continued on until the mid 19th Century. After the mid 19th Century, however, there was a major readvance which in some cases exceeded the initial late 17th to early 18th Century advance. Subsequent retreat has occurred, but many minor readvances and fluctuations in rates of recession have caused widespread formation of moraines. Recessional rates decreased in the early 20th Century, but again accelerated since 1930 A.D. and exceeded any previous rates attained in the last few centuries. Heusser compared the most recent glacial events with temperature and precipitation data recorded near Banff since 1895. He found that there was a lag in

the response of glaciers to climate in the order of about 30 years. However, Heusser also noted that the glaciers should have responded to an apparent climatic deterioration which began in the early 1940's; but after almost 40 years they have yet to respond (i.e., Luckman, 1977). This may indicate that the simple meteorological data used may be insufficient for interpreting glacial responses.

In addition, Heusser (1956) analysed a peat section near the Jasper townsite for arboreal pollen. This section consisted of sedge peat over limnic sediments in which a layer of volcanic ash was present. Heusser interpreted the pollen as indicating a generally cool postglacial period; followed by a slightly warmer and drier episode, contemporaneous with volcanism; which was, in turn, followed by a return to cooler climates, with the last few centuries being slightly more humid. Although Heusser's interpretation is simplistic by today's palynological standards and lacks absolute dates, it remains the only published source of pollen data within the Banff-Jasper area.

Luckman (1977) worked on the Holocene moraines at two sites in Jasper National Park and developed a chronology based on a lichen growth curve for *Rhizocarpon geographicum*. This growth curve was developed by using dendrochronological and historical evidence. Four of the most recent (Little Ice Age) moraines were dated at 1705 ± 5, 1720 ± 5, 1858 ± 7, and 1888 ± 7 A.D. at Mount Edith Cavell. Dates of 1765 ± 5, 1810 ± 5, 1876 ± 5, and 1907 ± 5 A.D. were obtained from four similar moraines at Penstock Creek, 13 km west of Mount Edith Cavell. Luckman estimated the recession of the Cavell

Glacier, once coalescent with the Angel Glacier, to be 16 m yr^{-1} from 1927-1963 A.D. and 6-8 m yr^{-1} from 1963-1975 A.D. The Angel Glacier has shown a similar pattern of retreat but has maintained its terminal position since 1962. This may be evidence of a glacial response to the post-1940 cooling (with increased precipitation) postulated by Heusser (1956), and would imply a lag time of 20 years for the Angel Glacier. However, the influence of topographic control on the terminal position of the Angel Glacier must not be precluded as this glacier presently ends at the foot of an icefall. Luckman also found Bridge River Ash (2600 BP, Nasmith, *et al.*, 1967) within the soils in front of the 1705 moraine at Mount Edith Cavell. This indicates that no greater ice advance has occurred for at least 2600 years. However, Luckman found older lateral moraine remnants along the valley sides at Mount Edith Cavell which were not overridden by the subsequent advances. There are at least three of these pre-Little Ice Age moraines and the minimum lichenometric age for the oldest is ~1660 BP. Evidence for earlier advances also occurs in the Banff area where McCallum and Wittenberg (1965) obtained a date of 6020 ± 90 for a soil overlain by ablation till.

Since most glaciers overrode evidence of early Neoglacial advances during the Little Ice Age, rock glaciers provide most of the geomorphic evidence for the earlier Holocene climatic fluctuations. Luckman and Crockett (1976) noted at least two phases of rock glacier development prior to the Little Ice Age. Some of these rock glaciers have also been reactivated or overridden by advancing ice glaciers. Surface lichens showed that these rock glaciers were equivalent to, or slightly older than, the oldest moraines at Mount Edith

Cavell. It appears that most of the present day rock glacier activity is limited to the "ice-cored" types of glacial origin (c.f. Potter, 1972). The majority of rock glaciers, however, are heavily lichen covered and lack any evidence for recent movement. From this evidence, Luckman and Crockett (1976) suggest that the major period of rock glacier development must have been prior to the Little Ice Age and possibly pre-Hypsithermal.

Osborn and Duford (1976) found Mazama Ash (6600 BP) on the surface of a rock glacier in northern Banff National Park. This provides good evidence that a climate favorable to rock glacier formation must have existed before the Hypsithermal. Mazama Ash was also found on a kame moraine adjacent to a Little Ice Age moraine of the Athabasca Glacier which also suggests an ice advance prior to the Hypsithermal. However, Osborn and Duford agree with the above workers that the Little Ice Age advances were the most extensive of the Holocene advances.

In summary, the evidence presented by the above workers suggests that there have been several phases of glacial readvance between Roed's (1975) Drystone Creek advance and the Little Ice Age. Although the Little Ice Age advances have been fairly well documented, the timing of the earlier advances remain speculative due to the absence of absolute dates and because the Little Ice Age advances were so extensive that most of the evidence for the earlier advances has been either destroyed, or confined to rock glaciers or sparsely scattered moraine remnants.

1.2.2 Western United States

Owing to the multitude of studies that deal with Holocene glaciations in the Cordilleran regions of the United States, only a few of the major works will be discussed here. Blackwelder (1915) defined three Pleistocene glaciations in the Wind River Mountains, Wyoming. This sequence has since become the type area for the Rocky Mountain Glacial succession. The three glaciations were called the Buffalo, Bull Lake, and Pinedale respectively. Blackwelder differentiated these advances largely on the relative amounts of subaerial weathering which has occurred on exposed surfaces. Deposits similar to these have since been widely recognized throughout the United States Cordillera. Richmond (1965) summarized the sequence of Rocky Mountain glaciations in the United States. He noted, from radiocarbon dates at the base of postglacial peat deposits, that the deglaciation of the Pinedale Advance in the lower altitudes was complete by ~7280 ± 400 BP (however, dates range from 11330 ± 330 to 8800 ± 250 BP), and not until ~6190 ± 300 BP in the higher altitudes. The soils of the Hypsithermal episode, in the northern Rocky Mountains of the United States, range from Chernozems in the lower altitudes to Brown and Brown Forest Soils, and locally to Podzols, at higher elevations. These soils in many cases have been overridden by deposits of the Neoglacial phase. Richmond also noted that, throughout western Montana, lenses of Mazama Ash occur in close association with these Hypsithermal soils, giving an approximate age for this intervall (6600 BP). Richmond (1962) recognized two minor advances in Rocky Mountain cirques since the end of the Hypsithermal. The older advance, the Temple Lake Stade (named after a moraine at Temple Peak in the southern Wind River Mountains),

has 6-18 m high moraines which are rough and bouldery, and covered with tundra vegetation, including small stands of scrub spruce. Most of these moraines are found from 1-3 km from the cirque headwalls and the till of this advance is fresh, sandy, and bouldery, with striated stones being rare. The second advance is called the Gannett Peak Stade (named after moraines near the terminus of the Gannett Glacier in the northern Wind River Mountains). These moraines are also fresh, rough, and bouldery, but they lack any soil or vegetation, except for a sparse lichen cover. The Gannett Peak moraines also lie upslope from the Temple Lake moraines and historic evidence shows that glaciers were at this position ~1850 A.D.

Benedict (1968) used a lichen growth curve for *Rhizocarpon geographicum* to date rock glaciers, moraines, protalus ramparts, bedrock surfaces, and talus slopes in the Indian Peaks region of the Colorado Front Range. The lichenometric dates obtained were partially verified by a few radiocarbon dates. Benedict found evidence of three major glacial episodes, each containing a series of minor pulsations. His earliest (Temple Lake) advance dated between 4500 and 2700 BP. The intermediate advance (Arikaree) were dated between 2100 and 1000 BP, and the most recent advance (Gannett Peak) at 1650-1850 A.D.

Miller (1969) used dendrochronology, lichenometry, and geomorphology to establish a scheme of Neoglaciation in the Dome Peak area, North Cascade Range, Washington. He dated the earliest, post Hypsithermal, advance at 4900 BP. However, Miller noted that most glaciers in the area reached their maximum extent sometime in the 17th

Century, and he did not find evidence for an intermediate advance.

Miller (1973) also determined a chronology of rock glacier development during the Neoglacial period in the Northern Sawatch Range, Colorado. He used lichenometry to subdivide the deposits into several age groups. Temple Lake I, the earliest period of rock glacier development, began >4000 BP and was followed by a period of decreased activity. After this time, a second stage of rock glacier development, Temple Lake II, began 3750-3500 BP and culminated at 2500 BP. Rock glaciers of intermediate age began forming ~1900 BP and continued to develop until 1000-900 BP during the Audubon period. The relative sizes of the Temple Lake and Audubon rock glaciers indicate that conditions favoring rock glacier development were of a higher amplitude, or of longer duration, during the Temple Lake interval. The most recent, Gannett Peak interval, on the other hand, is represented only by the formation of talus slopes rather than rock glaciers.

Mahaney (1973) used relative age dating techniques to develop a Neoglacial chronology at the Fourth of July Cirque in the central Colorado Front Range. Glacial and periglacial deposits were assigned ages based on a growth rate curve for *R. geographicum* which were then supplemented by additional relative age measurements pertaining to solum development, weathering rind thicknesses and pits on rocks, morphology of moraine crests, and the extent of patterned ground development. Three periods of glaciation were found. The earliest period was equated to the Temple Lake interval and dated

4500-2700 BP. The intermediate advance was equated to the Audubon interval and dated 950 BP (minimum estimate), and the youngest advance, the Gannett Peak, occurred at ~300-100 BP.

Benedict (1973) outlined the chronology of cirque glaciations for the Colorado Front Range, inferred by a number of radiometric dates. He found that there were at least 4 to 5 distinct intervals of glacial expansion during the Holocene. The earliest of these, the Santana Peak Advance, occurred shortly before 9915 ± 15 BP. With subsequent retreat, the timber-line rose to its present position by 9200 ± 135 BP, and by 8460 ± 140 BP patterned ground features on the Santana Peak moraines became inactive. There is also some evidence of a minor advance before 7900 ± 130 BP. Curry (1974) has radiometrically dated a moraine at 6500 ± 230 BP in Wyoming. Benedict found that the latter part of the Hypsithermal was characterized by the complete extinction of perennial snowbanks and cirque glaciers together with the formation of a soil. At 5000-3000 BP the Triple Lakes advance followed which was the most extensive Neoglacial advance. Evidence in the Arapaho Cirque suggests a three-fold sequence for this interval with advances dated at 4485 ± 100, 3865 ± 100, and ~3150 BP respectively. This glacial period ended with an interval of soil formation and weathering which, in turn, was followed by the onset of the Audubon Advance at 1850-950 BP. The Audubon was characterized by the formation of rock glaciers as well as cirque glaciers and perennial snowbanks above the treeline. Subsequently, active recession occurred at ~1505 ± 95 BP on the valley floors, and

~955 ± 95 BP for ice in the sheltered cirques. This recession was, in turn, followed by the most recent, Arapaho Peak Advance, occurring ~300-100 BP. Benedict chose to use local place names for his chronology in an effort to avoid premature correlation with poorly dated type localities and assumptions of apparent synchroneity.

1.2.3 Conclusion

Table 1.1 summarizes some of the chronologies discussed above. Although there appears to be some general agreement of inferred climatic trends for various sites throughout the North American Cordilleran, closer inspection shows that the exact timing of certain geomorphic events described is equivocal. This may be due to several factors. First, only a minimum age estimate could be obtained for a number of the events described. Furthermore, geomorphic events, such as glacial advances, may vary in response to regional climatic changes. The timing and extent of glacial responses from site to site is controlled by mesoclimatic, morphological, and lithological factors. The result is that glaciers in one valley may respond to an environmental change which has no effect in an adjacent valley, and suggests that correlation based on the stratigraphy of such events, without absolute dating control, is very tenuous. This may be especially true for the work done in the Canadian Rockies where some conflicting data exists.

Table 1.1 A comparison of some Rocky Mountain glacial chronologies.

$\times 10^3$ Years BP	Summary of U.S. Cordilleran Richmond (1965)	Indian Peaks Colo. Front Range Benedict (1968)	Dome Peak N. Cascade Range, Wash. Miller (1969)	N. Sawatch Range Colo. Miller (1973)	Fourth of July Cirque Colo. Front Range Mahaney (1973)	Colo. Front Range Benedict (1973)	Banff-Jasper National Parks Heusser (1950)	Jasper National Park Luckman (1977, 1978)	Banff-Yoho National Parks Osborn (1975) Osborn and Duford (1976)
0	NEOGLACIAL: Gannett Peak	Gannett Peak	Second advance	Gannett Peak	Gannett Peak	Arapaho Peak	Little Ice Age	Little Ice Age	Gannett Peak (> 180 BP)
1	NEOGLACIAL: Interstade	Arikaree		Audubon	Audubon (>950)	Audubon			Audubon (> 650 BP)
2								advance (> 1660 BP)	
3–4	NEOGLACIAL: Temple Lake	Temple Lake		Temple Lake II	Temple Lake	Triple Lakes			
4–5			first advance	Temple Lake I (>4000 BP)				?	
5–7	HYPSITHERMAL	Hypsithermal	Hypsithermal		Hypsithermal	Hypsithermal		Hypsithermal	Hypsithermal
7–9	PINEDALE: Late Stade					Minor Advance (>7900 BP)		? Divide deglaciated	advance (> 6600 BP)
10–11	PINEDALE: Inter Stade					Santana Peak (9915 BP)			

CHAPTER TWO

STUDY AREA: LOCATION AND DESCRIPTION

2.1 Location

The Whistlers Creek Valley (118°06'W and 52°49'N) lies approximately 11 km southwest of Jasper townsite (Fig. 2.1). Whistlers Creek flows in a northeasterly direction from the Continental Divide into the Athabasca River which comprises part of the eastern watershed of the Rocky Mountains. The creek is approximately 9 km long draining an area of ~22 km^2, and originates from the confluence of a number of streams which drain cirques at the valley head. The study area covers the upper portion of Whistlers Creek Valley (~5.5 km^2) and its location is noted in Fig. 2.1.

2.2 Description

A contour map of the study area is given in Fig. 2.2. Whistlers Creek Valley is surrounded by a number of prominent mountains which provide a variety of alpine features such as cirques, aretes, and horns. The valley separates The Whistlers and Marmot mountains which attain elevations of 2464 and 2708 m respectively (Fig. 2.2). In its upper reaches, the valley separates Terminal Mountain and Indian Ridge which have elevations of 2835 and 2743 m respectively. The head of the valley terminates at the foot of Manx Peak (elevation, 3096 m). Two well developed cirques, facing northeast; are currently occupied by small glaciers in the uppermost portion of the

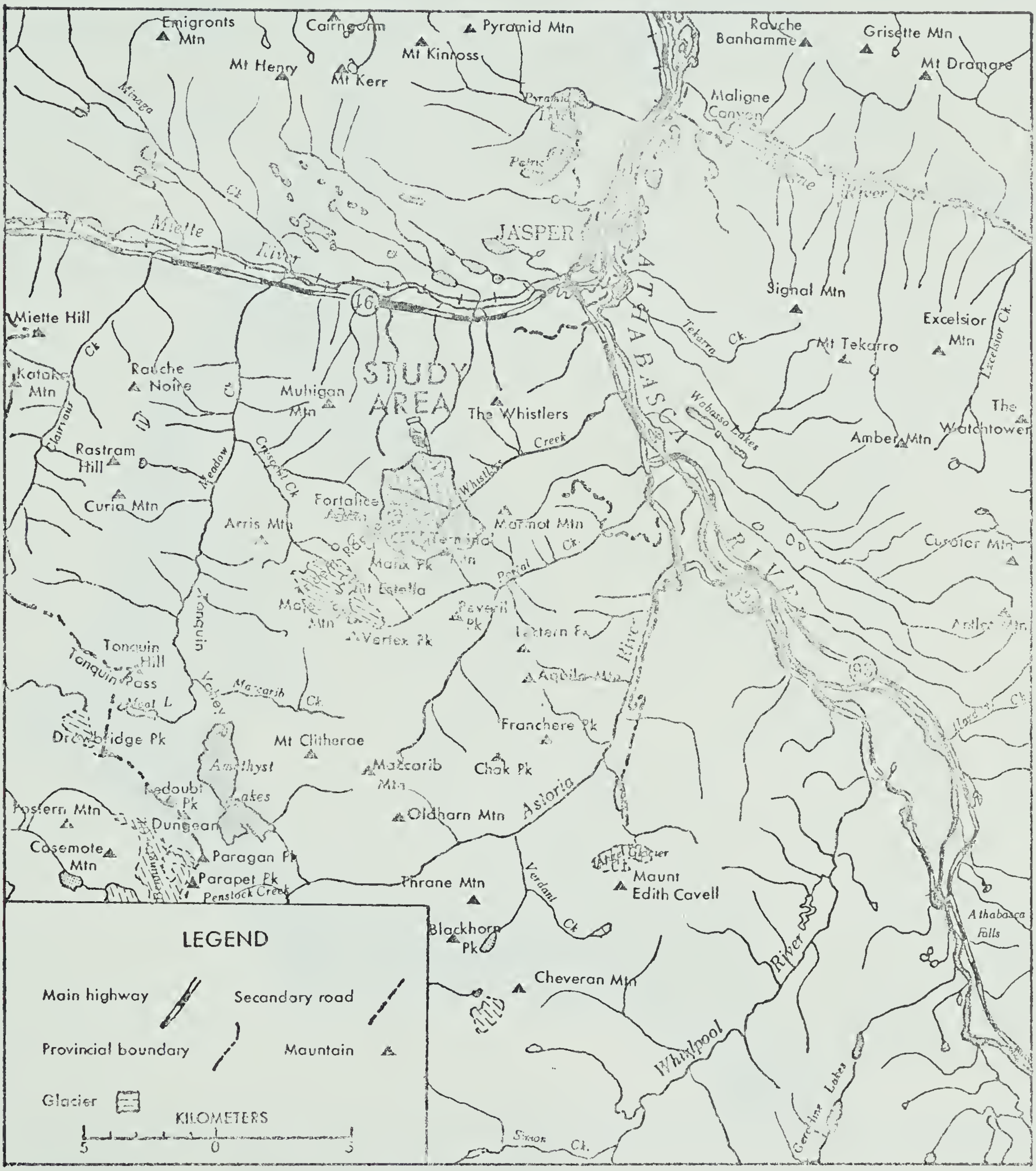


FIGURE 2.1 Location of the study area.

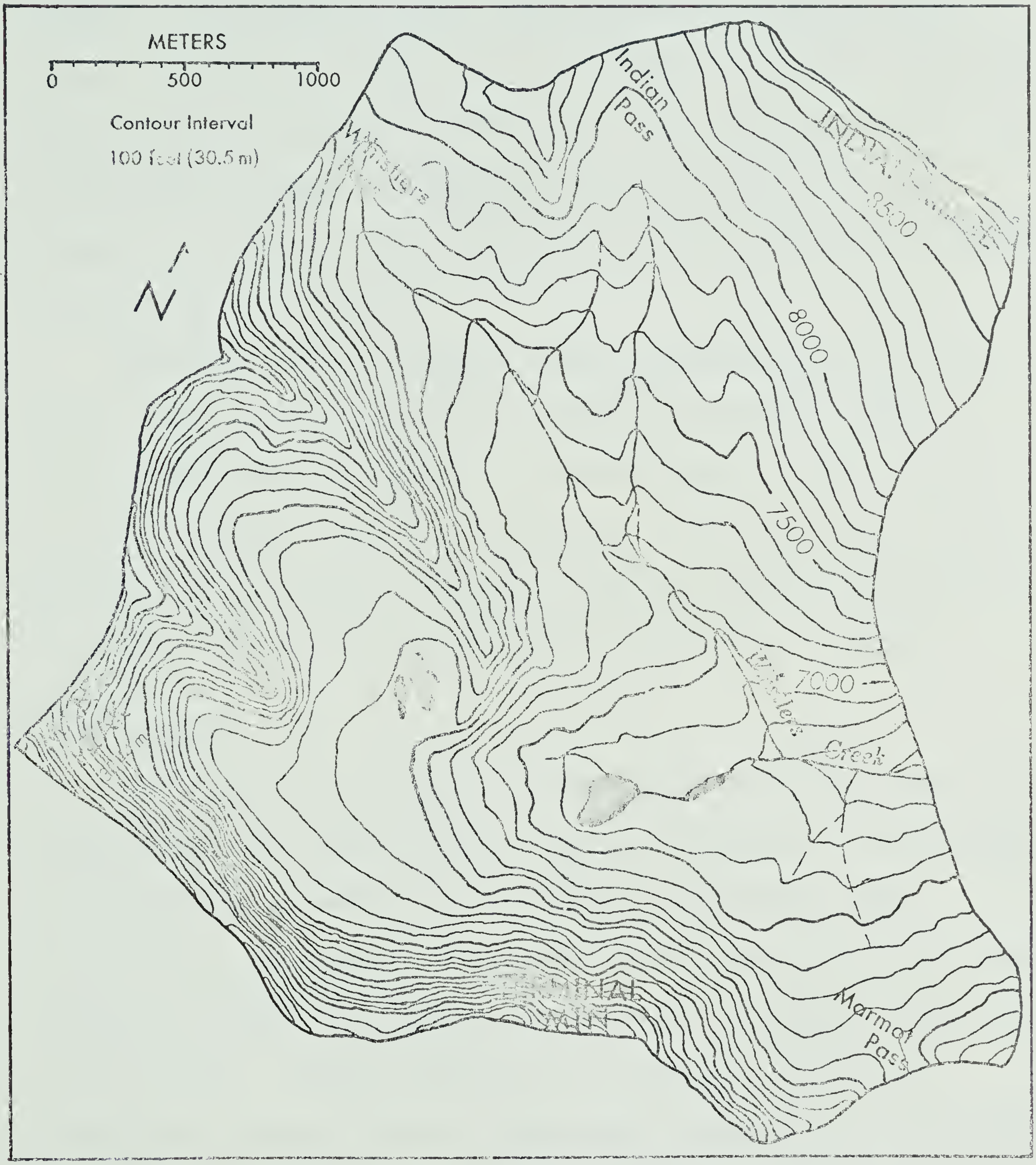


FIGURE 2.2 Contour map of the study area.

valley. The glaciers presently extend down to the 2290 m contour level. Perennial snow patches are also present in the more sheltered areas of Terminal Mountain and Manx Peak. The topography of Whistlers Creek Valley has been accentuated by extensive glaciation as the valley is a classical alpine trough which forms a hanging valley above the main Athabasca Valley.

The study area also exhibits a wide variety of periglacial features, particularly prominent rock glaciers and various types of solifluction lobes and terraces. Avalanche cones and talus slopes are also abundant, especially at the base of the steep cliffs formed by Manx Peak and Terminal Mountain. Detailed discussion of both the glacial and periglacial features will be undertaken in Chapter V.

2.3 Geology

The mountains surrounding Whistlers Creek Valley fall within the Main Ranges of the Rocky Mountains which are composed of folded shales and quartzites of Cambrian and Pre-Cambrian age (Baird, 1963). Specifically, the bedrock of the study area is composed of three major formations (Table 2.1) which strike southeast to northwest and dip to the southwest. The lower Cambrian, Gog Quartzite Formation is restricted to areas southwest of the Marmot and Whistlers Passes (Fig. 2.3). This formation is up to 1500 m in thickness and forms massive vertical cliffs (c.f. Mount Edith Cavell; Luckman and Crockett, 1978). Manx Peak and Terminal Mountain are good examples of this rock type, which tends to break up into large, roughly equidimensional blocks as

Table 2.1 The bedrock formations found within the study area (after Montjoy, 1977).

1.	Gog Formation	- Quartzose sandstone and shale (Lower Cambrian)
2.	Upper Miette	- Silty argillites, siltstones and thin bedded sandstones (Pre Cambrian)
3.	Middle Miette	- Fine conglomerates interbedded with argillite units (Pre Cambrian)

Table 2.2 Temperature estimates (^{o}C) for an elevation of 2300 m (from Janz *et al.*, 1977).

	January	April	July	October
mean maximum	- 8	0	+ 12	+ 2
mean free air	-11	- 7	+ 7	- 1
mean minimum	-14	- 8	+ 2	- 5

Height of July freezing level - 2700 m

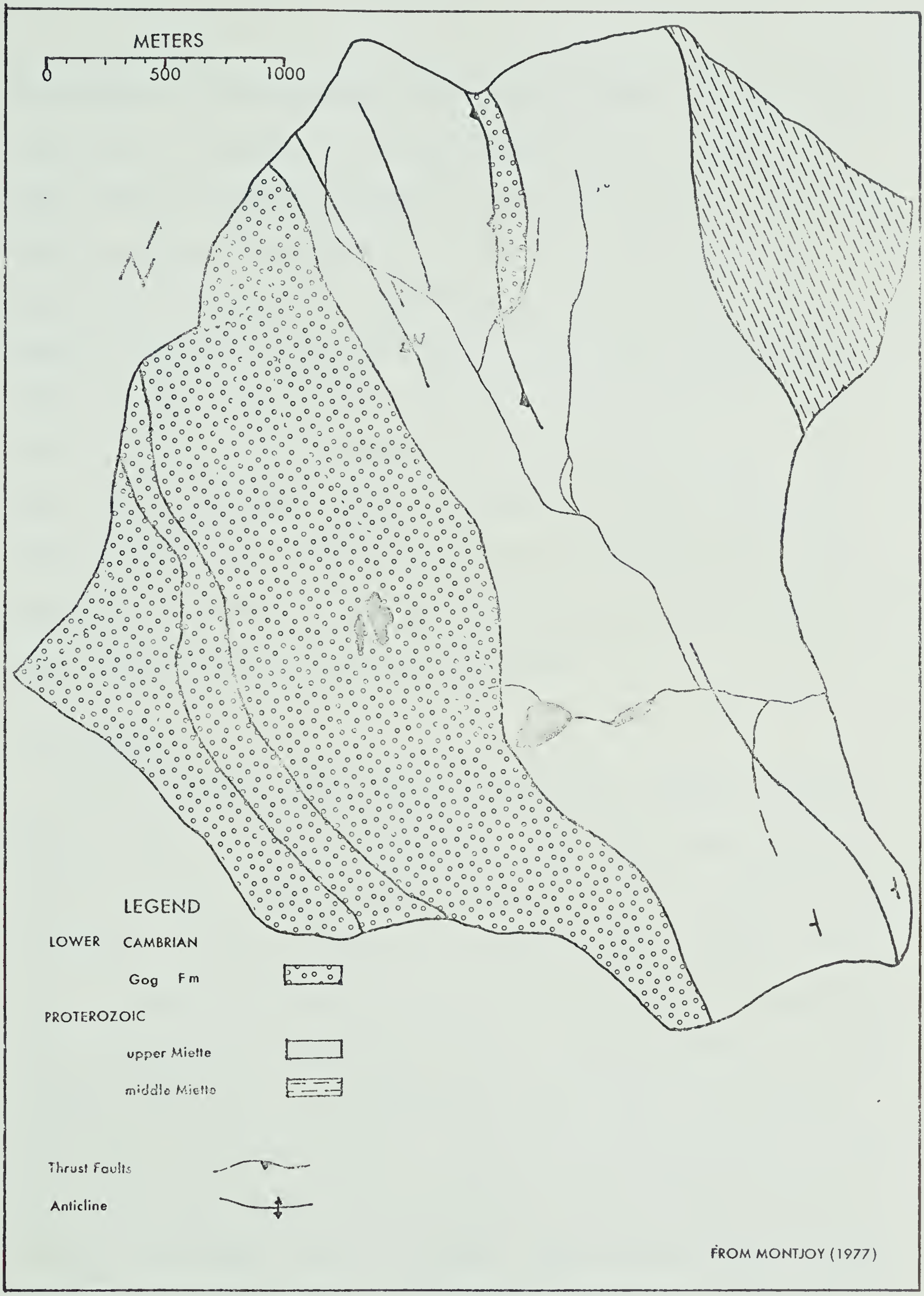


FIGURE 2.3 Bedrock geology of the study area.

it weathers. Stratigraphically below the Gog Group, the rest of the study area is composed of the Miette Group with the exception of a small amount of Gog being exposed on a minor peak between Indian and Whistlers Passes, the result of a thrust fault (Mountjoy, 1977). The Middle Miette Formation on Indian Ridge, southwest of the Whistlers Mountain, exhibits considerable folding and probably minor faulting (Mountjoy, pers. comm., 1977). Compared to the resistant Gog Formation, the Upper Miette Formation is weak and weathers easily. The major Passes, Whistlers, Indian, and Marmot lie within the Miette Formation (Fig. 2.3). Argillites and coarse grained quartzites and conglomerates within the Upper Miette Formation vary in resistance and form a number of tors within the study area.

2.4 Climate

Hare and Thomas (1974) refer to three basic climatic regions in Western Canada: 1) The Pacific Region, a maritime climate characterized by moderate temperatures and precipitation with a distinct winter maximum and a summer minimum, 2) the Prairie Region, having a continental climate characterized by large seasonal temperature fluctuations and a precipitation regime with a distinct summer maximum and a winter minimum, and lastly, 3) the Cordilleran Region which has influences from both of the former Regions. Storm tracks from the Pacific advect warm air and most of the moisture into the Rockies, while outbreaks of Arctic air masses from the northeast bring extreme minimum temperatures to the eastern parts of this Region. Temperature and precipitation regimes, then, are controlled by the succession of the two air

masses moving into the Cordilleran Region. Substantial seasonal variability can therefore occur, particularly due to changes in the upper Westerly circulation which tends to steer these surface air mass systems. The Westerly wind belt usually forms semi-stationary pressure troughs and ridges around the circumpolar area, commonly known as the Rossby waves (Rossby, 1939). The position, amplitude, and number of these standing waves then, ultimately influence the climate of the region.

Topography of the mountain sector also has great influences on the climate, especially on a local scale, making this region hard to generalize. Certain valleys may have a precipitation regime similar to the continental sector whereas higher elevations, on adjacent mountain sides, may experience frequent, orographically induced disturbances from the Pacific Region. The variations are due to a number of effects the mountains have on local climate which can be classed into two general categories: 1) the effects of elevation, and 2) the effects of aspect.

1) The general effects of increasing elevation are an increase in the percentage of solid precipitation, and a decrease in both the minimum and maximum temperatures with a corresponding increase in the total amount of annual freeze-thaw cycles. Table 2.2 shows the approximate annual temperature conditions for the Jasper area at an elevation of 2300 m (the average elevation for upper Whistlers Creek Valley). The values were derived from temperature-elevation relationships calculated by Janz and Storr (1977). Janz and Storr estimate that

derived temperatures should be within ± 2^{o}C, but the maximum could be as much as 3-5^{o}C lower in July due to the proximity of glaciers or snowfields. The amount of annual precipitation at the 2300 m level is estimated at 1050 mm (calculated from a regression derived by Storr and Ferguson, 1972).

2) Aspect, or the orientation and slope of a particular mountain, has several effects on the local climate. The major effect is to control the amount of insolation falling onto a specific surface. Measuring effective, short-wave radiation southeast of Banff National Park, Ferguson *et al.*, (1971) found that a south-facing slope may receive in excess of 800 calcm^{-2}da^{-1}, whereas the northeast-facing slope receives less than 500 calcm^{-2}da^{-1}. Geiger (1965) found that in Germany a south-facing slope of 20^{o} receives approximately twice the radiation that a horizontal surface receives in January. The orientation of the slope then, has a great effect on the local temperature regimes of mountain slopes and hence strongly influences the processes operating on them.

A second effect of aspect on the local mountain climate is that of the prevailing winds. The major effect of winds, in addition to reducing temperatures through wind chill, is related to the redistribution and spillover of snow which effects precipitation receipts. This is an important consideration because at an elevation of 2300 m ~75% of the total annual precipitation falls as snow (Janz and Storr, 1977). Spillover is a term used to describe snow which is blown from the windward side of a mountain, over the crest, and onto the lee side

where it accumulates. This phenomenon has the effect of greatly increasing the annual precipitation and snowcover on the lee sides of mountains.

The twofold effect of aspect affects the local climate of Whistlers Creek Valley in a number of ways. Firstly, the head of the valley is northeast facing which means that it receives the least amount of insolation, especially in the shaded areas at the base of the large cliffs composed of Gog quartzites. This condition, together with the reduced temperatures due to the higher elevation,means that sheltered patches of snow and ice are more likely to survive the summer ablation season. Secondly, the northwest to southeast orientation of the mountain ranges puts the continental divide at virtually 90° to the prevailing westerly winds with the head of Whistlers Creek Valley forming an effective catchment area for the spillover of solid precipitation. The result is that the amount of actual snowcover predicted for this area is greatly increased,and this factor significantly contributes to the annual increment of accumulation on glaciers and snow patches found in the study area. In addition, abundant moisture supply from snowmelt greatly aids the solifluction processes discussed in Chapter IV.

The number of freeze-thaw cycles in the study area is an important climatic factor when considering periglacial processes as it relates to slope movement, sorting, and mechanical weathering. Janz and Storr (1977) suggest that in mountainous areas the maximum number of freeze-thaw cycles (with mean temperatures several degrees below

freezing and mean maximum temperatures several degrees above freezing) would occur in midsummer at approximately the 3000 m level. Fraser (1959) tentatively classed the number of freeze-thaw cycles in the Jasper region between 60-80. This figure may,however,vary considerably with the changing topographic factors discussed above.

A second factor thought to be important in the periglacial environment is the presence of permafrost or perennially frozen gound. Péwé (1969) considers all areas underlain by permafrost to be in the modern periglacial zone. Reports of permafrost in the Rocky Mountains of Alberta are meager. Ogilvie and Baptie (1967) first described permafrost at Snow Creek Valley, Banff National Park (Fig. 1.1). At an elevation of 2347 m, the mean annual temperature recorded over a 2 year period was -3.9^{o}C, and thaw occurred to a depth of 38 cm by September. Scotter (1975) reported ice lenses at a depth of 130-210 cm and permafrost below 210 cm on Lookout Mountain, Banff National Park. This occurrence is on a steep southwest-facing slope at an elevation of 2655-2691 m. Based on these observations, comparable elevations and temperature estimates for Whistlers Creek Valley suggest that the study area should contain permafrost, especially on northeast-facing slopes.

2.5 Vegetation

The vegetation within the study area is only briefly dealt with here. About 90 percent of the study area lies above the tree-line (Fig. 2.2). The tree-line in Whistlers Creek Valley varies somewhat

with changing microenvironmental conditions, but it may be generalized at ~2200 m. The forests below tree-line consist mainly of Alpine fir (*Abies lasiocarpa*) with isolated Englemann spruce (*Picea englemanni*) which tend to be very old (~300 years). The vegetation at the tree-line becomes characteristically krummholz with dwarf species of birch and willow being common. As the shrub vegetation terminates, heaths such as Mountain heather (*Cassiope mertensiana*) and Rocky Mountain white heather (*Cassiope tetragona*) dominate. At higher elevations the vegetation becomes progressively depauperate until only lichens and bryophytes grow.

2.6 Place Names

Most of the place names used in this study are found on the Jasper, 83 D/16, East half, NTS, 1:50000 map sheet. Some additional names in the text are used for convenience and are unofficial. All names used are indicated on the Place Name Map (Fig. 2.4).

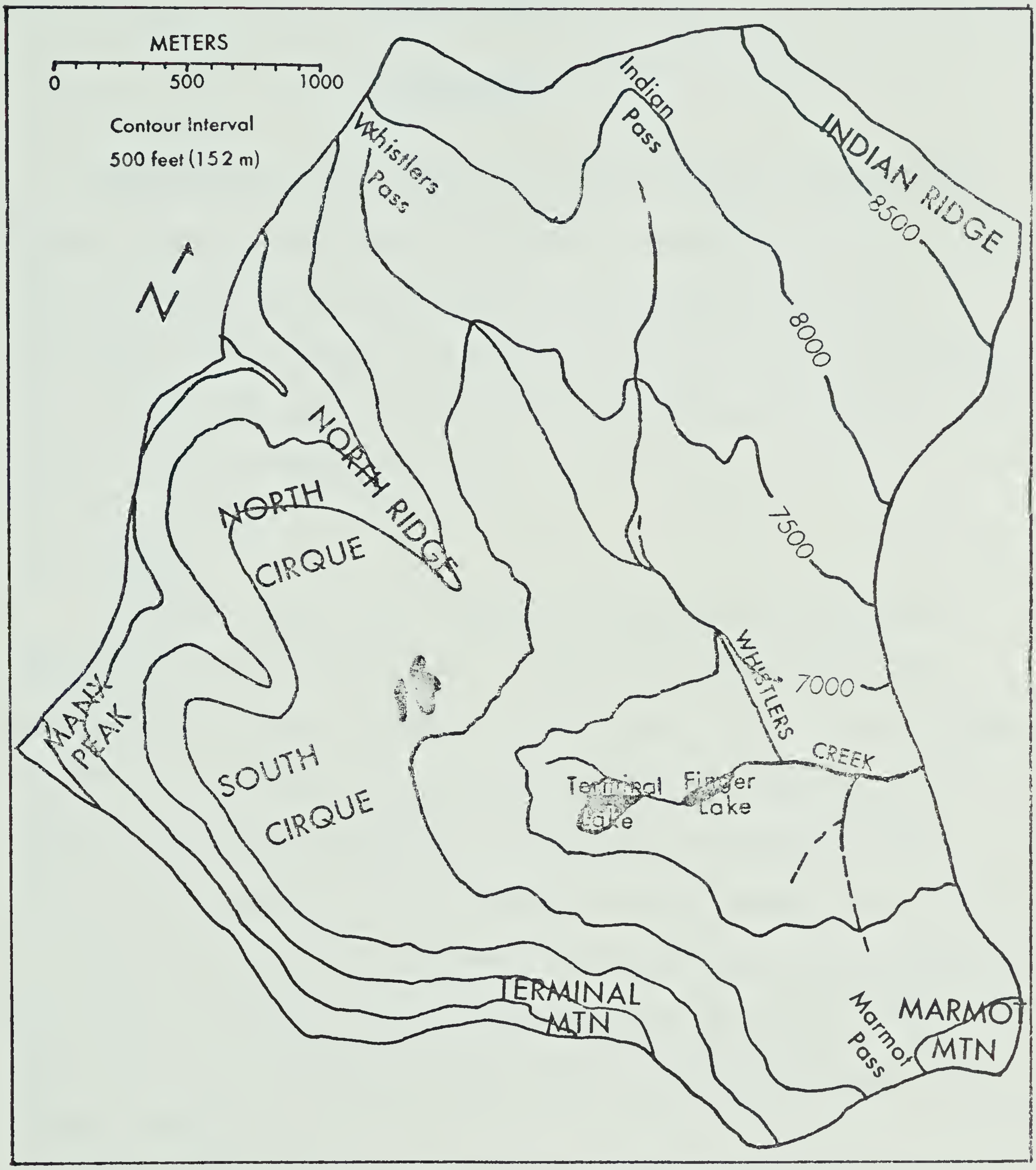


FIGURE 2.4 Place names .

CHAPTER THREE

METHODOLOGY

The methodology used in this study was concerned with two objectives. First, to describe the landforms and surficial deposits within Whistlers Creek Valley, and secondly, to determine the history of geomorphic activity since the recession of the last, late Pleistocene ice advance. The research procedures were divided into mapping, field observations, and laboratory work.

3.1 Mapping

Preliminary maps of the landforms and surficial deposits were obtained from aerial photographs. Three sets of photographs were used for this purpose: A12407-81 to 82, taken in 1950, scale 1:40,000; A16085-1 to 10, taken in 1958, scale 1:12,000; and A23015-67 to 68, taken in 1972, scale 1:80,000. All three sets of photographs were taken in late summer. This photographic coverage spans 12 years which consequently provided an opportunity to consider short-term changes in the alpine landscape. The air photographs were also used to make approximate measurements on areal extent and gross dimensions of the larger landforms.

The preliminary air photograph interpretation was checked and further refined in the field and the resulting maps are given in the next chapter on landforms. In order to determine elevational differ-

ences between the mapped units, the air photographs were superimposed onto the 1:50,000 NTS map using a zoom transfer scope. However, because of the extreme relief in the valley, there was too much distortion in the air photographs for the depicted landforms to be precisely adjusted to fit the contours of the map. Due to this difficulty the resulting map only gives the approximate locations of the landforms and deposits. This problem is further compounded by the close spacing of contour lines found on maps of mountainous terrain which means that small errors in the location of mapped units may represent substantial jumps in elevation. The preceding difficulties then, limit the accuracy of the interpretive map in Chapter IV.

3.2 Field Study

The field work consisted of ground-proofing the air photograph interpretations, describing the observed landforms and surficial deposits, and estimating their relative ages.

3.2.1 Verification

Verification of the preliminary maps was accomplished by field mapping. Each mapped unit was visited and studied. In addition, the units were photographed and sketched from vantage points.

3.2.2 Description

The landforms were described with two main considerations in mind; firstly, to determine the morphology of the feature in question, and second, to note any evidence of geomorphic processes associated with the feature. The description of the features consisted of the following:

A. Measurement of Size and Slope

The smaller features, whose precise dimensions could not be obtained from air photographs, were measured on the ground by tape, whereas their slopes were determined by Abney level. Transects were surveyed on some of the larger landforms (such as the rock glaciers) in order to obtain cross-sectional profiles.

B. Excavation

Where possible, the landforms were partially excavated and any evidence of pebble orientation and sorting was noted. Further samples were taken for laboratory analysis. However, the massive nature of the debris in this valley often made excavations impractical.

In addition to the above, the morphology of the landforms was considered in a dynamic framework as well. Any evidence of differential movement or overriding within the larger landforms (rock glaciers) was considered both in respect to their process of formation and to their chronology of activity. The various relative age dating techniques discussed below, also served to further describe the features.

3.2.3 Relative Age Dating

Several techniques have been successfully used for relative age determinations of glacial events. A selection of studies employing a number of these techniques have been mentioned in Chapter I. Most relative age dating techniques depend on phenomena which alter rock surfaces exposed to the subaerial environment since glacial action terminated. The duration of this exposure, or the length of time the

surface has been stable, is what is being measured. The following relative age dating techniques have been used in this study:

A. Topographic Position

Each feature's relationship to adjacent landforms was noted. Initial age determinations were based on cross-cutting relationships and superposition of the landforms. When considering glacial deposits, the older units are generally higher on the valley walls and further downstream than younger deposits. This is because as a younger advance occurs it usually destroys the older deposits it overrides (cf. Luckman and Crockett, 1978). Clearly, however, the time difference between such features must be determined by more detailed observations as outlined below.

B. Geomorphic Character

Geomorphic character refers to the general appearance and expression (morphology) of the feature. The criteria considered were: first, measuring the sharpness of ridge crests and steepness of slope angles. Ridge crests tend to become rounded and slope angles decrease as the landscape becomes more subdued with increasing age (Miller, 1973; Kiver, 1974). Second, to note the stability of the surface. For example, assess whether the surface is loose and easily disturbed versus stable and vegetated.

C. Lichenometry

Lichenometry proved to be most useful for understanding the formation and chronology of activity of the rock glaciers in Whistlers Creek Valley. Beschel (1950, 1961) pioneered the study of

lichens as a chronological tool. The technique is based on the observation that certain epipetric (crustose) lichens quickly colonize a freshly exposed surface and grow at a determinable rate for very long periods of time. Consequently, lichen growth is a function of time and prevailing microenvironmental conditions. Given a constant environment then, the relative age of the surface will be reflected by; the lichen thalli diameters, the species succession (Orwin, 1970), and the percentage of the total surface covered by lichens. However, the problem of assuming a constant environment is made obvious by the fact that one is measuring the age of geomorphic features which owe their very presence to climatic change. Consequently, paleoclimatic change makes a precise growth rate curve questionable.

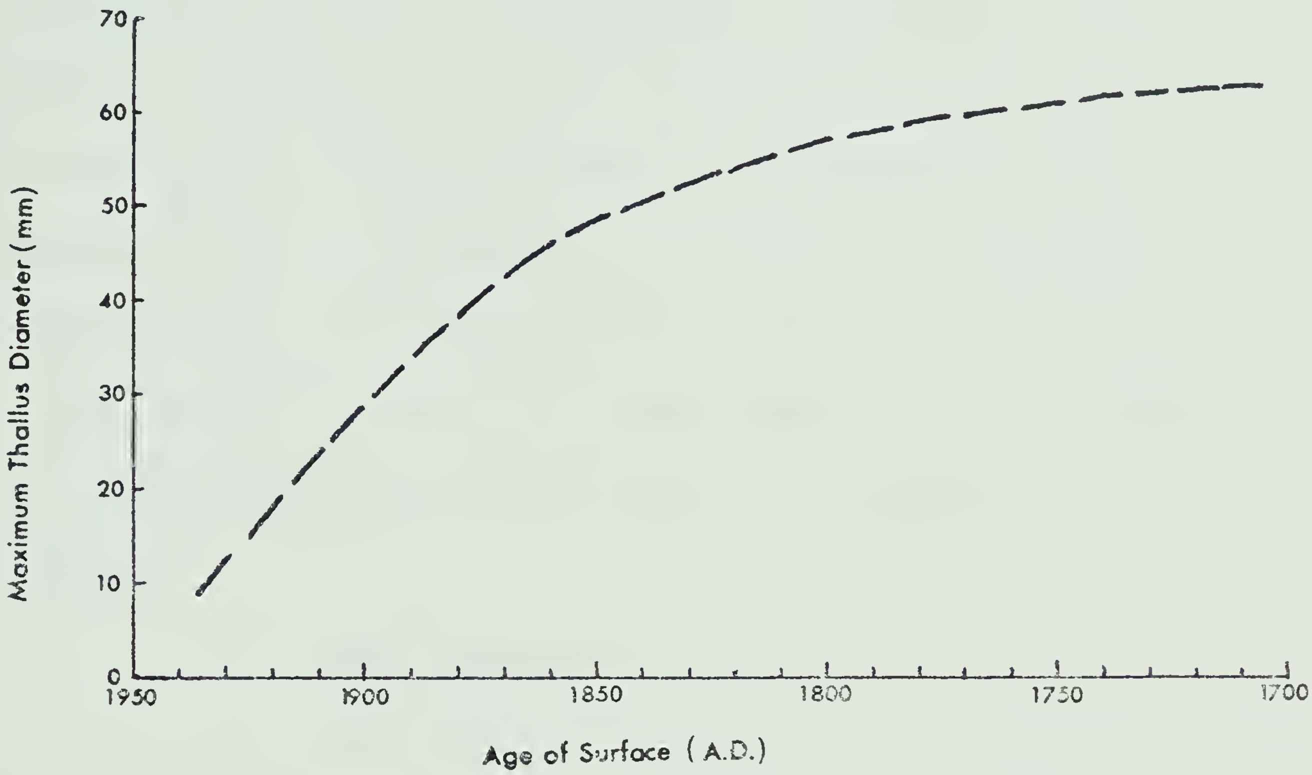
Several species of lichen have been used for dating geomorphic surfaces. *Rhizocarpon geographicum* is most commonly used, however, because of its longevity, circular growth, and widespread geographical distribution. *R. geographicum* prefers to grow on siliceous rocks and is therefore frequently found on the quartzites within Whistlers Creek Valley.

Beschel (1950) initially reasoned that if the actual growth rate of a particular lichen species could be determined, the absolute age of the substrate could also be derived by simply measuring the size of the thallus. The problem is, however, that the growth rates vary from site to site as microenvironmental conditions change. Therefore, in order for the determined growth rate to be applicable to one's area, the microenvironmental influence must be eliminated.

Beschel reasoned that this factor could be eliminated by comparing only the largest thalli between sites. By doing so, he thought he would be comparing only the optimal environments at each site. These optimal conditions then, were thought to represent similar microenvironments between sites.

Subsequent studies have found Beschel's assumption to be true, but only for sites of close geographical location and similar setting. Webber and Andrews (1973), and Jochimsen (1973) have recently reviewed this problem in greater detail. A growing number of studies have successfully utilized lichenometry to interpret Neoglacial events (cf. Benedict, 1967, 1968; Denton and Karlén, 1973; Miller, 1973; Osborn, 1975; Luckman, 1977).

Luckman (1977) has developed a lichen growth curve for *R. geographicum* in the Mount Edith Cavell area of Jasper National Park (Fig. 3.1). This preliminary growth curve spans a 250 year period. After an inception period of approximately 30 years, the lichens appear to grow 42 mm/100 yrs. for the first 110 years,and then 11.4 mm/100 yrs. for the subsequent 140 years. Since Mount Edith Cavell is only 10 km south of Whistlers Creek Valley it seems likely that the microenvironments are similar (Luckman, Pers. Comm., 1977), hence some absolute dates were determined in Whistlers Creek Valley using Luckman's growth curve. In addition, three specimens of *Xanthoria elegans* were measured in the study area and their thalli were compared with a growth curve developed by Osborn and Taylor (1975) for the Banff/Jasper area.



Growth curve for *Rhizocarpon geographicum* in the Cavell-Tonquin area.

FIGURE 3.1 Luckman's (1977) growth curve.

R. geographicum was identified in the field by its color and pattern. Criteria for identification were acquired by studying similar specimens at the University of Alberta Herbarium. In addition, representative samples were collected in the field and later identified with laboratory methods by M. Ostafichuk, University of Alberta. Thalli diameters were measured to the nearest millimeter and the percentage of surface covered by all species was determined by visually comparing the surface in question with coverage charts compiled by Terry and Chilingar (1955). Determination of species successions was not attempted in this study. Further discussion of the problems encountered with lichenometry in the study is given in Chapter V.

D. Rock Weathering

The weathered condition of surface boulders in glacial deposits has long been used as a determinant of relative age (cf. Blackwelder, 1931), and this technique has aided a number of glacial studies in the Cordilleran region (Richmond, 1965; Madole, 1969; Mahaney, 1972, 1973, 1974, 1975; Birkeland, 1973; Kiver, 1974). Surface weathering is usually evaluated by a number of different methods: 1) by the depth of rind penetration (oxidation) on exposed boulders (Nelson, 1954; Williams, 1973; Porter, 1975); 2) by the angularity of surface rocks (Caine, 1968; Birkeland, 1973; Kiver, 1974); 3) by the extent of rock surface roughening (crystal projection) and pitting (Birman, 1964; Dyke, 1976); 4) the size reduction of boulders, and their progressive burial by grus, has also been incorporated into

weathering studies (Birman, 1964; Sharp, 1969). The procedures used to determine rock weathering in the study area were as follows:

a) Determinations of rind thickness were made by fracturing surface stones with a rock hammer and then measuring the depth of discoloration to the nearest millimeter. Five or more stones were measured at each site along extensive transects, and the average was then determined.

b) The angularity of rocks at each site was determined by comparing the shape of the rocks in question with photographs of clay models depicting six degrees of roundness (Powers, 1973). The roundness classes used were: very angular, angular, subangular, subrounded, rounded, and well rounded.

c) The extent of weathering and pitting was determined by measuring the relief of inclusions and individual grains, or crystals, on the rock surface, and by measuring the depth of weathering pits.

It must be noted here that all the rock weathering measurements were confined to similar looking quartzites (i.e. similar texture). In addition, areas where anomalous weathering rates may occur (i.e. near water) were avoided. By the above methods extraneous factors influencing weathering rates, other than age, were minimized. Discussion of the problems using these techniques in the study is given in Chapter V.

E. Soils

The extent of soil development is another useful way

of distinguishing surface age on the basis of weathering. Soils have been used extensively by Benedict (1970), Mahaney (1972, 1974, 1976), and Birkeland (1973, 1974) for dating Neoglacial events in the U.S. Cordillera. Soil profiles, and many of their prominent properties, require hundreds to thousands of years to form, hence, soil morphology and clay mineralogy are two main characteristics which can be used for relative age dating. The use of soils for relative age dating, however, has many restrictions, especially in a highly variable alpine environment. Problems of local variability may be minimized by obtaining a large sample size, but for logistical reasons this was not attempted in this study. Soils were, therefore, of a secondary importance to this study. Three soil pits were dug and their locations (Fig. 3.2) were selected to detect differences on surfaces of predetermined relative age. Therefore, the locations of the soil pits were chosen after the relative ages of the surfaces had been tentatively determined by the other dating techniques. It must be stressed, however, that the results are only suggestive because of the small sample size and the highly variable nature of alpine microenvironments.

The morphological criteria used in this study were:

a) The development of the A horizon organic matter. Organic matter within soils is assumed to reach a steady state in a relatively short period of time, with estimates ranging from 200-300 years (Birkeland, 1974). However, if the surfaces studied are fairly young, the progressively older soils should show increasing organic content and dispersal patterns with respect to

time.

b) The development towards a textural B horizon. The clay distribution in soils of similar parent material is a useful criterion for determining relative age of a soil. Clay content usually increases in amount and thickness in the B horizon with time.

c) Clay minerals can also be used for determining relative age. When considering clay minerals, some basic generalizations must first be made. Firstly, clay minerals are in equilibrium with the environment in which they have been formed, otherwise they would not exist. Secondly, for parent materials with low initial clay content, there is little evidence that clay minerals vary with time (Birkeland, 1969; Birkeland and Janda, 1971). Thirdly, if clays are initially present in the parent material, some of the assemblages may be unstable and will gradually change into more stable forms. The objective then, is to determine the initial clay assemblages (C horizons) and then look for alterations that may have taken place in the solum. The initial assemblages must, however, be similar for each of the soils considered.

d) Soil pH. There is a tendency for the pH, within the soil, to decrease with increasing age (Mahaney, 1974; Birkeland, 1974).

The three soil pits were dug and described according to the Canadian System of Soil Classification (1977). The soil

horizons were further analysed for percent organic matter, particle size, clay minerals, and pH. The procedures used for these analyses are described in the section on Laboratory Work.

F. Dendrochronology

Measuring the ages of trees growing on moraines and related features has long been used as a method of minimum dating (Heusser, 1950; Lawrence, 1946, 1950; Mathews, 1953; Miller, 1969; Luckman, 1977). Stokes and Smiley (1968) describe this technique in detail. In order to obtain a better estimate on the minimum age of a certain depositional feature, the inception period for trees colonizing that newly exposed surface must be determined. For the Mount Edith Cavell area, Heusser (1956) estimated an inception period of 12-17 years. However, Luckman (1977) estimated the inception period for the same area to be from 15 years to 30 years in the more xeric sites. Chronological investigations have also been made on the tilting of trees along trimlines by advancing ice (Heusser, 1950). When a tree is tilted by some external force it responds by growing reaction wood within its annual layers (Scurfield, 1973). It is possible then, to determine when a tree has been disturbed by examining its growth rings.

The 16 trees sampled in this study were drilled with a Swedish increment corer, and the 5 mm diameter cores were mounted on grooved boards and sanded until the annual rings were easily distinguishable. The rings were then counted under a telescoping microscope and the presence of reaction wood was recorded. All the trees sampled were either Alpine fir or Englemann spruce. The approximate location

of the trees cored is shown in Figure 3.2.

G. Other Techniques

Tephrachronology and radiocarbon dating, which may have provided time-stratigraphic relationships between Whistlers Creek Valley and other locations, could not be applied. Although volcanic ash layers have been found on surficial deposits in various areas of the Alberta Cordillera (Heusser, 1950; Westgate and Dreimanis, 1967; Osborn and Duford, 1976; Westgate, 1977), no evidence of ash has yet been found in Whistlers Creek Valley. An ash-like substance was found in soil pit 2 (S-2, Fig. 3.2), however, petrographic analysis failed to reveal any traces of volcanic ash (Rutter, Pers. Comm., 1978). In addition, no relevant material for radiocarbon dating was found, although extensive traverses and excavations were made. The apparent absence of both volcanic ash and preserved organic debris in the upper Whistlers Creek Valley is probably the result of its high energy, periglacial, surface environment, and the lack of suitable catchment areas. However, there may be potential for finding preserved organic debris and ash on the bottoms of the deep, rock basin lakes (see Chapter IV).

3.3 Laboratory Work

The laboratory work consisted of analysing the samples collected in the field for: particle size distribution, organic content, clay minerals, and soil pH. Figure 3.2 shows the locations where the samples were taken.

3.3.1 Particle Size Analysis

Particle size analysis was used to describe the samples

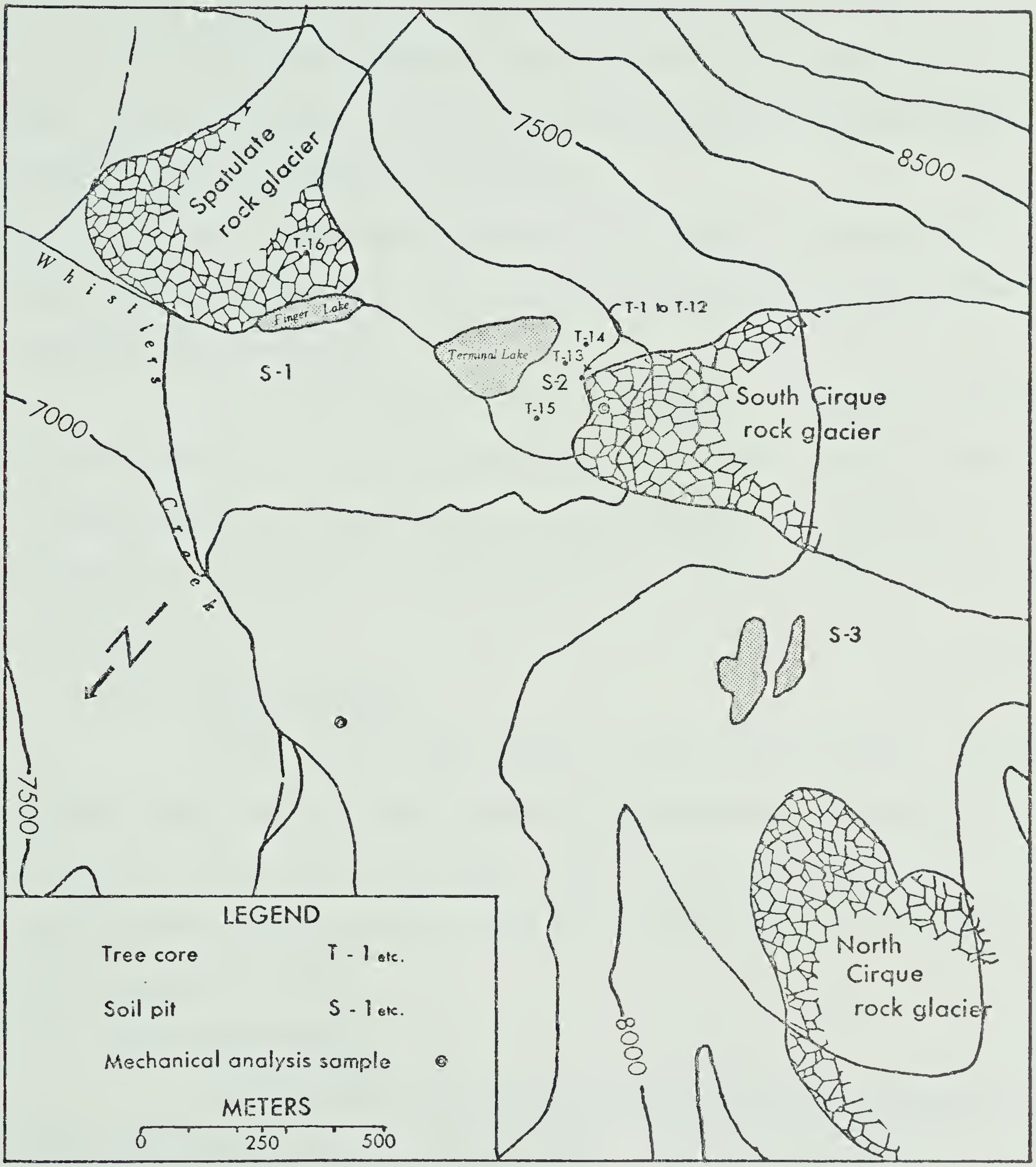


FIGURE 3.2 Location of the sample sites.

taken (i.e., the soil profiles).

The method of particle size analysis consisted of:

a) Dry Sieving the <2 mm size fraction through the -0.5, 0.0, 1.0, 2.0, 2.5 and 3.25 phi sized sieves, and wet sieving through the 4.0 phi sieve (sand/silt boundary).

b) Fractionating the silt into sizes of approximately 4.6, 4.9, 5.4, 6.0, and 6.4 phi units by an accelerated sedimentation method using a Cyclosizer.

c) Determining the silt/clay boundary (9.0 phi) by weighing the amount of clay extracted through sedimentation (Black, 1965). In some samples, in which clays were not extracted, the silt/clay boundary was determined by interpolating the cumulative grain size curve to 9.0 phi.

3.3.2 Organic Carbon

Carbon is the chief element of soil organic matter. Since organic carbon can be readily measured quantitatively, it was used as a means of estimating total organic matter within the soil. The Walkley-Black method (Allison, 1965) was used to extract the organic carbon from the samples.

3.3.3 Clay Minerals

The clay sized fraction was extracted from the samples by the sedimentation technique (Black, 1965). The clays were then centrifuged and washed to remove all salts, after which, each sample was split in two and part was Ca saturated and part was K saturated. The samples were then mounted on slides using the paste technique

(McKeague, 1976) and analysis of the clay minerals was carried out by X-ray diffraction.

3.3.4 Soil pH

Soil pH was measured by the standard method in 0.01 M $CaCl_2$, outlined by Peech (1965).

CHAPTER FOUR

MORPHOLOGY AND DISTRIBUTION OF THE LANDFORMS AND SURFICIAL DEPOSITS

This chapter contains a description of the landforms and surficial deposits mapped in the Whistlers Creek Valley. The location of the landforms in the study area and an air photo mosaic are shown in Figure 4.1, and Plate 4.1 respectively. Particular emphasis is placed on rock glaciers as these are the most outstanding geomorphic features in the study area. The geomorphic units identified and their associated landforms and surficial deposits are as follows:

4.1 Bedrock Features

Whistlers Creek Valley contains a wide variety of bedrock, topographic features commonly associated with glaciated, alpine landscapes. Two well-developed cirques lie at the foot of Manx Peak (Fig. 4.1). These are referred to as the North and South Cirques (Fig. 2.4). The South Cirque lies immediately below the headwall of Manx Peak which rises 450 m above the cirque floor. The walls surrounding this cirque contain numerous topographic irregularities which harbour permanent snow and ice. To the north, this cirque is bordered by an arete (North Ridge) which separates it from the North Cirque. The bedrock walls surrounding the North Cirque are about 200 m high. Additional glaciated bedrock features in the field area include cols, horns, riegels, spurs, and the 'U' shaped, hanging valley of Whistlers Creek itself.

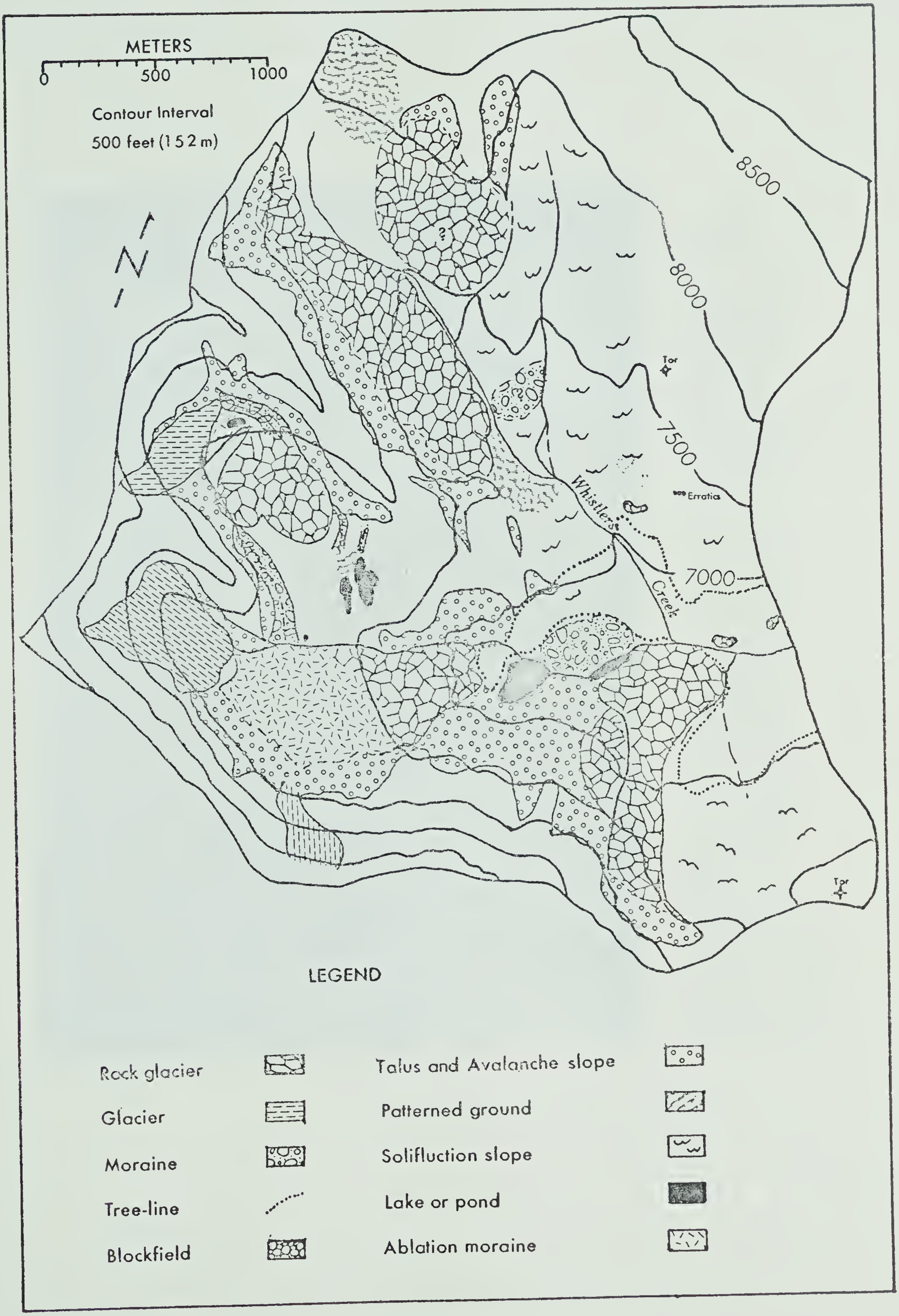


FIGURE 4.1 The Landforms in the study area .

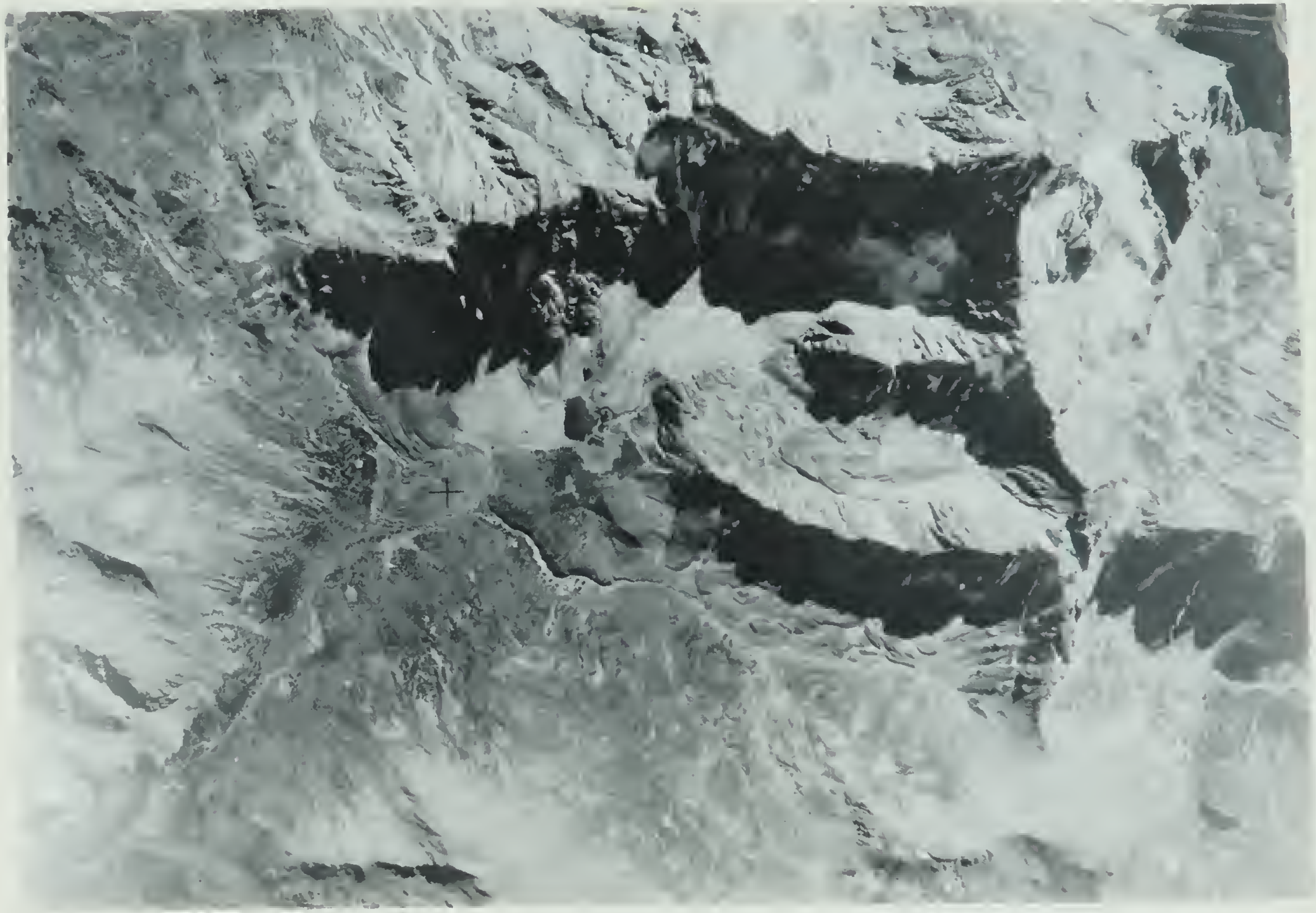


Plate 4.1 An air photograph of the study area. (North is toward the bottom.) Note the spatulate rock glacier (left of center) and the North and South Cirques with their respective glaciers and rock glaciers. Heavy lichen cover on the spatulate rock glacier is indicated by its dark tone. (Dept. of Energy, Mines and Resources, Ottawa, air photo: A12407-82)

In addition to the glacially altered bedrock features, there are a number of landforms associated with the interbeds of resistant and non-resistant rocks of the Upper Miette Formation. This structural control has resulted in the formation of large scale bedrock terraces. These terraces are most evident on the north side of Whistlers Creek Valley, running the length of Indian Ridge (Plate 4.2), where they form a series of large steps. The formation of these features may have been accentuated by the ice that occupied this valley in the past (see Sec. 4.4.2).

The more resistant beds of the Upper Miette Formation have also led to the formation of several tors within the study area. The best examples of these are on the west flank of Marmot Mountain (Plate 4.3). These tors are composed of coarse grained quartzites and conglomerates with veins of white quartz, and rise about 3-4 m above the surrounding surface. Their appearance is very weathered, as extensive frost shattering has occurred and some fractures extend throughout the entire tor. The average weathering rind thickness was 6 mm and the largest *R. geographicum* thallus was 130 mm, however, the thallus was weathered out near its center. The quartz veins running through the rock protrude 3-5 mm from the surface.

Several tors of low relief (~1 m) also occur on the south slope of Indian Ridge. These tors are found on the edges of some of the large bedrock terraces described above (Plate 4.2), and are comprised of the more resistant conglomerates of the Upper Miette Formation. These tors are well-rounded in appearance and a large amount of

Plate 4.2 Bedrock terraces on Indian Ridge. The view is towards the north. Prominent solifluction terraces are also visible. Whistlers Creek flows along the base of the slope.

Plate 4.3 A tor on the west flank of Marmot Mountain. Extensive frost shattering is evident, with fractured debris scattered around its periphery.

granular disintegration has taken place, infilling the joints and covering the surrounding ground with debris. Lichens are absent from their surfaces which may be due to the advanced weathering observed. The differences in morphology between these tors and the tors on Marmot Mountain seems to be largely the result of the susceptibility to weathering of the constituent rocks.

4.2 Water

Surface water is widespread in the study area and its presence is likely enhanced by permafrost above 2300-2600 m (see Sec. 2.4). It is found in the form of streams, lakes, and numerous ponds. In addition, it is noteworthy that many of the solifluction slopes remained saturated throughout the 1977 summer. The most prominent stream in the area, exits from the South Cirque rock glacier, and flows into Terminal Lake. As this lake is dammed by a moraine, it must drain beneath the surface, through coarse debris, into Finger Lake. This lake in turn, drains into the main valley channel which receives additional inflow from many streamlets in Indian and Whistlers Passes (Fig. 2.4 and 4.1).

A small spring also drains out of the snout of the Spatulate rock glacier and flows into the main channel. The only other lakes found within the study area, occur in the North Cirque, and occupy deep depressions in the cirque floor which are probably the result of glacial scour. In addition, numerous ponds and streams are found on the surface of the rock glaciers and incised into the ablation moraine.

Perennial snowbanks and some forms of patterned ground also favor local surface water.

4.3 Ice and Permanent Snow

Two small glaciers occupy the cirques at the head of the Whistlers Creek Valley (North and South Cirques, Fig. 4.1). These glaciers are wholly contained within the cirques and fall into the "cirque glacier" category (cf. Flint, 1971, p. 28).

4.3.1 South Cirque Glacier

The cirque glacier occupying the South Cirque has an area of ~0.2 km^2. At present, the terminus of this glacier lies at an elevation of 2360 m, and the riser part of a riegel (slope 25^{o}) lies exposed immediately downslope (Fig. 4.1). Transverse crevasses, indicating extending flow approximately 50 m behind the terminus, may indicate the location of the crest of the riegel. The terminus of the glacier itself is very steep (35^{o}), and during the 1977 summer, considerable ablation till was being exposed and transported by mass wastage down the front of the glacier. In late summer the local snowline occurred at an elevation of ~2440 m. The abundance of debris on the ablation zone of this glacier suggests that avalanching plays an important role in its overall annual accumulation. In addition, many avalanche couloirs in the headwall lead onto the surface of this glacier, and although the winter frequency of avalanching is not known, avalanches occurred frequently during the 1977 field season. As many as three occurrences a day were noted, which brought snow and debris

onto the surface of the glacier.

A secondary source of ice in the South Cirque is a small "niche glacier" (cf. Embleton and King, 1975, p. 95) situated in a crevice between Manx Peak and Terminal Mountain (Fig. 4.1). No observations could be made on this glacier because of its inaccessibility.

4.3.2 North Cirque Glacier

A second cirque glacier (~0.1 km^2) occupies the North Cirque (Fig. 4.1). It occurs at approximately the same elevation as the South Cirque glacier, suggesting similar long-term snowlines for both localities. Much less debris is visible on the surface of this glacier than the North Cirque glacier, perhaps due to its smaller headwall which reduces the frequency of avalanching and rockfalls. Seventy-five meters downvalley from this glacier, a bedrock wall extends about two-thirds the way across the cirque floor. This wall has been extensively moulded and polished by ice, and, in the past, must have partially dammed the glacier. This feature will be discussed further in the rock glacier section (4.5.1). At present, there is a large depression between the bedrock wall and the terminus of the glacier. However, the 1950 air photographs show no such depression, suggesting that during the past 27 years, this glacier has experienced a negative net mass balance. This response is similar to other observations on glaciers in Western North America (Mier and Post, 1962), however, it is stressed that current mass balance studies are not available for these local glaciers.

4.3.3 Permanent Snowbanks

In addition to the glaciers,there are numerous perennial snowbanks found within the valley. The majority of these occur in the crevices and niches on the northeast-facing cliffs. Their presence is significant both for headwall destruction by nival processes, and for debris supply to the cirque floors. There are also a number of semi-permanent snowbanks in the valley which survived the 1977 ablation season. These are also visible on the 1972 and 1958 air photographs. However, they are absent on the 1950 air photographs. Since the more climatically sensitive perennial snowbanks have persisted through the last few decades, this suggests that the post-1950 recession of the North Cirque glacier may reflect a lag response to pre-1950 climatic conditions.

4.4 Till

The till in Whistlers Creek Valley occurs both as moraines and thin drift sheets which blanket the lower parts of the valley floor.

4.4.1 Moraines

The only distinct lateral and terminal moraine systems in the study area are those associated with the expansion of the glacier in the North Cirque. Lateral moraine-like ridges occur in the South Cirque,but these are thought to be partially a product of the rock glacier and will be discussed under that section. An ablation moraine grades into the rock glacier occupying the South Cirque,and two moraine-like landforms are found on the valley floor.

A. Lateral and Terminal Moraines

A prominent lateral moraine occurs on the north slope of the North Cirque (Fig. 4.1). It is situated immediately above the present glacier and runs downvalley for ~500 m until it appears to grade into an adjacent rock glacier (Fig. 4.2). The moraine has a sharp crest which lies about 10 m from the cirque wall. The constituent material is fresh, angular and lying at an angle of repose of 37^{o}. The lichen cover ranges from zero, near the head of the glacier, to 5% where it grades into the rock glacier. Here, the maximum measured thallus is 38 mm. Immediately below, and running parallel to this moraine, is another smaller lateral moraine which has a very fresh appearance and is devoid of lichen cover. This lateral moraine appears to grade downvalley into the uppermost lobe of the aforementioned rock glacier (Fig. 4.2). The relationship of these moraines to the rock glacier will be discussed in Section 4.5. Further downvalley, near the front of the rock glacier in the North Cirque, lies a remnant terminal moraine (Fig. 4.2). This moraine has only 2-3 m of relief and moderate, stable slopes of ~21^{o}. The crest of the moraine is rounded and interstitial fines infill the entire surface. The surface boulders are subangular to angular and have more than 50% lichen cover. Individual lichen thalli were hard to distinguish since most of the lichens have coalesced. The largest thallus found was 65 mm. Approximately 50 m further downvalley from the above moraine, another terminal moraine lies aligned with a bedrock ridge on the cirque floor. However, the classification of this feature as a moraine is

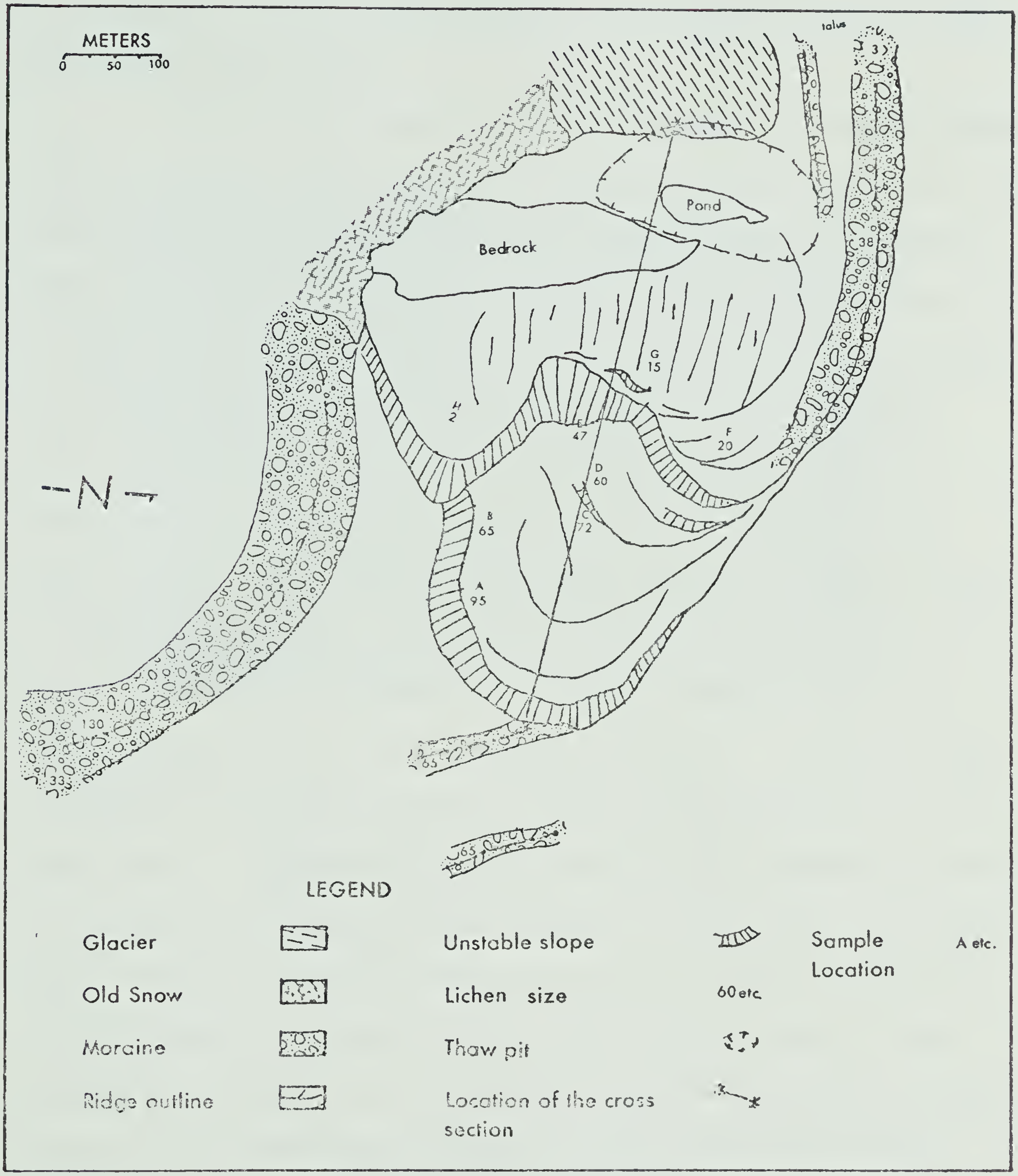


FIGURE 4.2 Plan view of the North Cirque rock glacier .

only tentative as it is poorly exposed and may only be a bedrock ridge with a thin mantle of till.

The largest moraine in the North Cirque is a lateral moraine situated along the south side of the cirque (Plate 4.4, Fig. 4.2). This moraine is also the furthest downvalley of the moraines in the North Cirque, and occurs at the highest position above the cirque floor. It is approximately 400 m in length and rises over 30 m from the surrounding terrain, and its sides have a slope of 36°. It is composed of large quartzose blocks, ~3 m across (long axis), which appear to be stacked on one another as no interstitial fines are visible in the upper section. The blocks are subangular to angular in shape and have a lichen cover of >50%. The largest *R. geographicum* found was 130 mm, however, its thallus was eroded at the center. The majority of the larger lichens appeared to be similarily weathered. The average rind thickness on the blocks is 2-3 mm. As this lateral moraine exits the North Cirque, it appears to have been truncated by a subsequent advance of the South Cirque glacier which occurred at some unknown time in the past. In this locality the slope of the lateral moraine steepens to ~40° and lichens are absent on many of the quartzite blocks. Those which do have lichens are 0-10% covered and the maximum thallus found was 33 mm. The constituent material in this sector of the moraine decreases in size, with blocks ranging from 30-60 cm in length, amongst some interstitial fines. The slope is unstable and it appears that some of the larger blocks in the upper section of the moraine have tumbled down and lodged at the base of the moraine.

Plate 4.4 A pre-Little Ice Age lateral moraine in the North Cirque. (North is to the right.) The North Cirque rock glacier is visible in the background, the South Cirque rock glacier-ablation moraine in the foreground.

There are two additional landforms in the study area which may be end moraines. These features are very subdued and covered by turf and shrubs. The evidence that they are moraines is based on their positioning, morphology, and composition (a mixture of unconsolidated fragments of coarse and fine debris, with no evidence of underlying bedrock). One of these tentative moraines dams Terminal Lake and the other is found near the Lobate rock glaciers (Fig. 4.1).

B. Ablation Moraine

The term "ablation moraine" is used here to describe a thin veneer of debris covering glacier ice (cf. Flint, 1971). Embleton and King (1975) note that an ablation moraine forms as englacial debris accumulates on the surface of a downwasting glacier.

The feature classed as an ablation moraine in this study lies between the riegel and the rock glacier in the South Cirque (Fig. 4.3). This moraine is considered to represent the transition between a debris choked glacier and a rock glacier. The principal difference exists in the ratio of ice to rock debris comprising these various bodies. Whereas the ablation moraine has a high concentration of ice and a relatively thin mantle of debris, the rock glacier is dominated by its debris cover (see Sec. 4.5). The thermal regimes of the underlying ice change as the debris cover increases. Østrem (1959) has stressed the efficacy of a debris cover in thermally preserving the underlying ice. Foster and Holmes (1965) have called a similar ablation moraine complex a "transitional rock glacier". Because this ablation moraine represents a transition, the boundary between it and

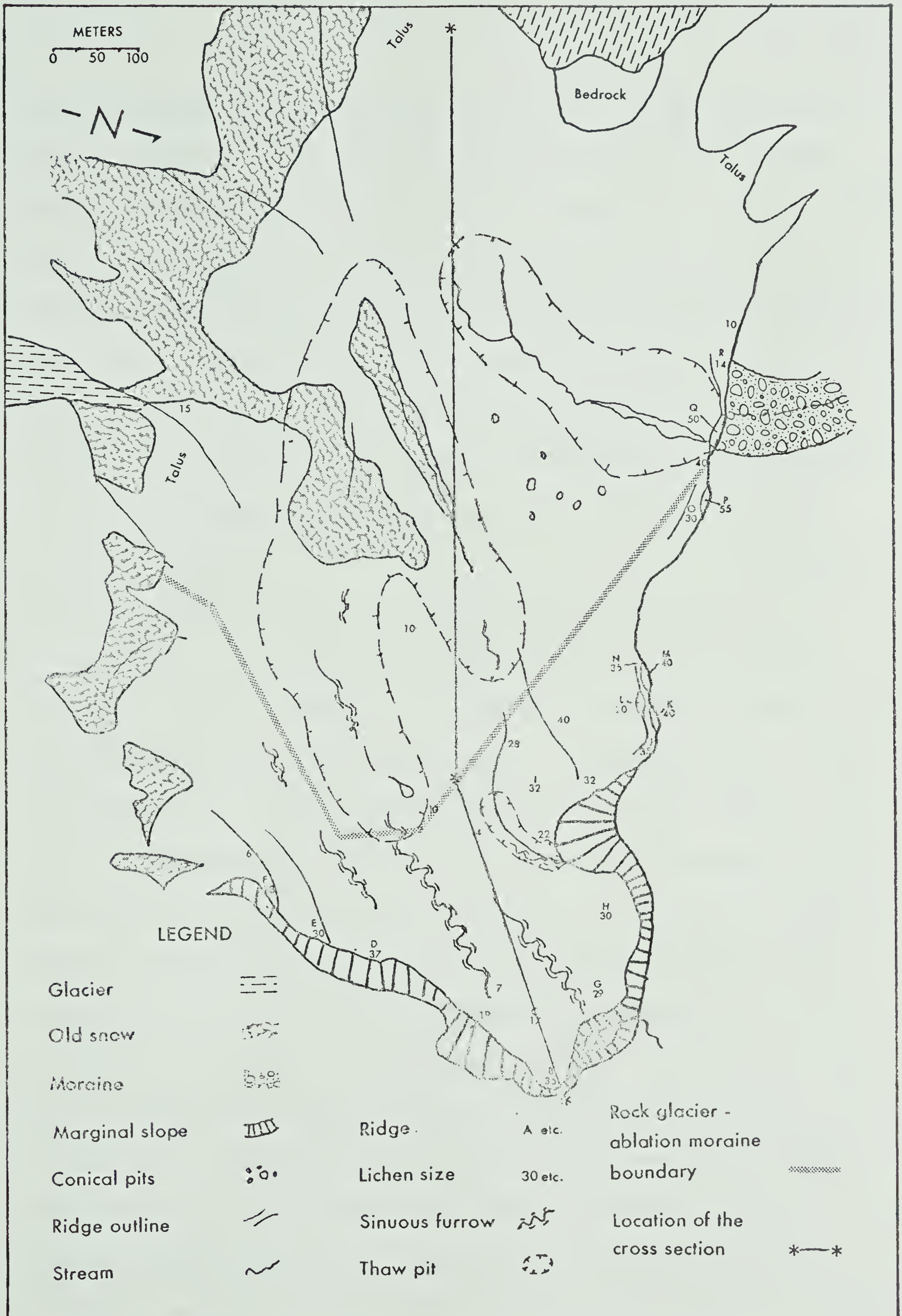


FIGURE 4.3 Plan view of the South Cirque rock glacier - ablation moraine.

the rock glacier is arbitrarily placed (Fig. 4.3). The boundary is based on the fact that on the ablation moraine there are more fines visible near the surface, and the overburden material is somewhat smaller, than the rock glacier. In addition, where the ice-rock interface was visible on the ablation moraine, the fines were concentrated immediately above the ice, whereas there were no exposures of clear ice on the rock glacier. The upvalley sections of the ablation moraine showed ~8 cm of fines overlying clear ice. This ice is evidently of glacial origin as it has foliations and debris bands dipping ~25° upvalley, and in some cases, crystals of ~3.5 cm in length are visible. Rock fragments found within the ice were also aligned with the foliations.

The surface morphology of the ablation moraine is complex and varied. Surface relief occurs in the form of large rounded ridges running the length of the moraine. These longitudinal ridges are separated by large fosse and depressions up to 25 m deep and they can be traced upvalley to debris cones at the base of adjacent cliffs. Inspection of the previous winter's snowcover on these cones showed an abundance of debris incorporated into it. This would suggest that avalanches and rockfalls must have played a significant role in the formation of the ridges on the ablation moraine. As the glacier was being covered by the above processes, it must have acted like a conveyor belt, transporting the debris from the base of the cliffs, downvalley. The relief between the ridges was probably further enhanced through differential melting of the underlying ice as it stagnated, leading to the development of the present ablation moraine .

Surface streams on this moraine have also penetrated the thin debris layer and have become incised into the ice. These streams commonly disappear into moulins after flowing on the surface for a few tens of meters. Large cavernous holes are also exposed on the surface of the ablation moraine. These holes are thought to be sections of large, relatively horizontal tunnels within the ice which have subsequently been exposed as the ice mass downwasted (Plate 4.5). As this moraine is related to the South Cirque rock glacier, further discussion on it will be dealt with in Section 4.5.

4.4.2 Drift Sheets

The second form of till in the study area is found as a thin and discontinuous drift sheet. Evidence for this drift sheet occurs in exposures along a gorge formed by Whistlers Creek. The bedrock of the gorge is composed of grey shales and siltstones which contrast markedly with the reddish quartzite diamicton mantling the surface. This material has a very coarse texture, comprised of large blocks and sands with negligible amounts of interstitial fine silts and clays. The textural composition of this diamicton, and the fact that there are erratics found within the same area, suggest that it is an ablation till (c.f. Flint, 1971).

The best example of erratics in the study area are found as a number of very large isolated blocks found on the north slope of Whistlers Creek Valley (Fig. 4.1). The blocks are massive pieces of greenish quartzite, the largest measuring 4 m in height and 6 m in

Plate 4.5 Cavernous openings on the South Cirque ablation moraine. Note the thin veneer of fines overlying glacial ice.

length. These blocks are considered to be erratics since they are of a different lithology than the surrounding bedrock (Upper Miette Formation). In addition, inspection of the bedrock outcroppings above these blocks also indicate that they did not originate from slopes at higher elevations. The nearest comparable bedrock is found across the valley, but deposition by a rockslide is precluded because of the isolated nature of these blocks, their extreme elevation, and the absence of others in the large, intervening area. The most probable conclusion, therefore, is that they were transported to their present position by glacial action. Similar greenish quartzite blocks were encountered on the summit of Whistlers Pass 2.5 km to the west (Fig. 4.1). In this area lichens were absent except on the lee (east) side of the blocks. The largest *R. geographicum* found was 75 mm and the average rind penetration was 7 mm. Transport of these quartzite blocks eastward from Whistlers Pass appears to indicate diffluent ice flow from a glacier occupying the main Miette Valley which entered into this interior valley system from the west. The age of this proposed advance is unknown. Such diffluent flow would suggest a minimum estimate of 1200 m of ice in the Miette Valley. The extent to which the topography of Whistlers Creek Valley has been altered by such "outside" ice, as opposed to the local cirque glaciers, is not known. Perhaps, the bedrock steps on Indian Ridge, mentioned in Section 4.1, were accentuated by this ice.

4.5 Rock Glaciers

4.5.1 Introduction

Although, strictly speaking, rock glaciers are a deposit of slope debris (Section 4.7), they are by far the most dominant landforms found within the study area, and hence, they are considered separately and in greatest detail.

Capps (1910) first used the term "rock glacier" to describe landforms of a specific surface morphology in Alaska. The definition of a rock glacier, used in this study, is also morphological and is well described by Potter (1972, p. 3027) as:

> "... a tongue-like or lobate body usually of angular boulders that resembles a small glacier, generally occurs in high mountainous terrain, and usually has ridges, furrows and sometimes lobes on its surface, and has a steep front at the angle of repose ..."

This morphologic definition was used since a review of the literature made it apparent that rock glaciers could be the product of a number of geomorphic processes and that on one genetic definition could satisfactorily explain them all. Comprehensive reviews and discussions of rock glaciers are given by Wahrhaftig and Cox (1959), White (1971), and Potter (1972). Rock glaciers have been explained as: special kinds of glacial moraines (Kohn, 1900; Kesseli, 1941); products of large landslides (Howe, 1909; Patton, 1910; Mudge, 1965); ice glaciers buried by debris (Cross and Howe, 1905; Brown, 1925; White, 1971; Potter, 1972;

Barsch, 1973; Whalley, 1974), or as a result of a flowing mixture of ice and rock (Spencer, 1900; Capps, 1910; Tyrell, 1910; Hole, 1912; Roots, 1954; Wahrhaftig and Cox, 1959; Thompson, 1962). Other processes suggested for rock glacier motion include; solifluction of interstitial mud and clay (Andersson, 1906; Chaix, 1923); talus creep (Kesseli, 1941), talus sliding over snowbanks (Parsons, 1939); and by freeze-thaw activity (Capps, 1910; Tyrell, 1910). Although some of the aforementioned processes may contribute to rock glacier movement, recent calculations based on rock glacier thickness versus surface slope (Wahrhaftig and Cox, 1959) have shown that basal shear stresses in active rock glaciers exceed the capability of frost creep and solifluction alone.

It is apparent that the theories of rock glacier formation are as numerous as the people writing about them. Currently, the two most favored models explaining rock glaciers are the flowing mixture of ice and rock in the interstitial ice model, and the debris covered, glacial ice-core model. The interstitial ice model (cf. Wahrhaftig and Cox, 1959) suggests that a mixture of ice and rock reaches a critical thickness and deforms under the influence of gravity. The debris then, acquires its flow-like structure from the resulting movement. In this view rock glaciers form when scree accumulates at the base of a cliff and ice develops in its interstices. This ice may be derived from avalanching (Thompson, 1962), refreezing of meltwater from adjacent glaciers (Capps, 1910), or by the freezing of emerging groundwater (Tyrell, 1910). In order for this ice to form,the mean

annual temperature within the debris must be less than $0^{o}C$. This condition depends on the local environment, particularly the insulation offered by the top few meters of debris, and by Balch ventilation (Thompson, 1957; Barsch, 1973).

The second favored model for rock glacier development suggests that rock glaciers are formed when a glacier becomes completely covered with detritus (Whalley, 1974). The glacier may be covered up catastrophically, as in a landslide (White, 1971), or gradually by prolonged downwasting at its surface which becomes buried by ablation moraine (Outcalt and Benedict, 1965; Benedict, 1973). Any subsequent movement then is due to the flow of the buried ice core together with the related shifting of the overlying debris mantle. This model, therefore, suggests that there is a gradation from clean ice glaciers, to debris covered glaciers (ice-cored rock glaciers), to stagnant rock glaciers containing no ice.

Although controversy still exists as to how rock glaciers form (i.e., Barsch, 1971; Potter, 1972; Whalley, 1974; White, 1976), many authors recognize that rock glaciers may form as a result of combinations of different geomorphic processes. Potter (1972) distinguished between two genetic classes of rock glaciers. These two classes fall within a continuum ranging from clean glacier ice to debris with essentially little or no ice. These classes are "ice-cored" rock glaciers, mainly representing buried glacier tongues, and "ice-cemented" rock glaciers, representing the interstitial ice theory. Outcalt and Benedict (1965) distinguished between "cirque-floor" rock

glaciers (ice-cored) and "valley-wall" rock glaciers (interstitial ice). Wahrhaftig and Cox (1959) proposed a morphological scheme for rock glacier classification. Their types are based primarily on the plan view of the rock glacier and are as follows; "lobate", in which the length of the rock glacier is less than the width, "tongue-shaped", in which the length is greater than the width, and "spatulate" which is a tongue-shaped rock glacier with an increase in the width at the front.

Perhaps the most interesting features of the rock glaciers are the ridges that form on their surface which give the rock glaciers their flow-like appearance. These ridges may be longitudinal or transverse and arcuate in shape, and are associated with furrows which are generally 'V' shaped in cross section. The longitudinal furrows are generally straight but occasionally have highly sinuous forms. Other forms of microrelief found on rock glaciers include conical pits, large depressions at the heads of the rock glaciers, and multiple lobes.

Wahrhaftig and Cox (1959) have extensively discussed the various forms of microrelief found on rock glaciers. Longitudinal ridges and furrows are thought to form when ice-rich bands melt out within a rock glacier. These ice-rich bands develop when snow accumulates in the swales found between adjacent talus cones feeding the head of the rock glacier. Subsequent downvalley movement of the rock glacier elongates the ice-rich swale areas to form bands. The furrows may be further deepened by meltwater thermally eroding into the ice. Should a furrow be deepened sufficiently, the debris on the sides of

adjacent ridges would be forced to readjust and rest at the angle of repose. This would result in a furrow with a 'V' shaped cross section. Some longitudinal ridges were also thought to form when there are differential rates of flow across a rock glacier. This may be especially true for some of the prominant lateral moraine-like ridges found on some rock glaciers. A decrease in the amount of ice along the sides of a rock glacier may lead to an increase in viscosity of the material which causes shearing along the margin of the rock glacier. Wahrhaftig and Cox (1959) considered that transverse ridges and furrows were the result of wrinkling of the non-plastic surface layer, or of internal shearing resulting in overthrusting, as the motion of a rock glacier slows in the front. Outcalt and Benedict (1965), on the other hand, felt that all ridges on a rock glacier were formed by successive readvances of an ice glacier. The rock glacier then, was composed of a growing complex of end moraine ridges. Potter (1972) felt that transverse ridges might result from glacial crevasses being infilled with debris coupled with subsequent melting of the ice which caused a reversal of relief. He also thought they may be flow features similar to the annual folds (ogives) beneath an icefall (cf. Raymond and Kamb, 1967).

Conical pits may result from the collapse of debris into vertical tunnels within the ice-rich parts of a rock glacier (Wahrhaftig and Cox, 1959; Potter, 1972). Sinuous furrows are thought to be caused when a series of closely spaced conical pits coalesce. Such furrows may also be analogous to the highly sinous pattern acquired

by meltwater streams on ice glaciers (cf. Baird, 1966, p. 91).

Lobes on rock glaciers may indicate multiple phases of development (Wahrhaftig and Cox, 1959) caused by renewed activity on the rock glacier after a period of quiescence. The renewed activity is thought to start at the head of the rock glacier where a new one forms and travels down the surface of the pre-existing rock glacier. The large pits at the heads of some rock glaciers may have been occupied by ice glaciers in the recent past (Wahrhaftig and Cox, 1959), or they may be the result of the melting of a buried ice core (Outcalt and Benedict, 1965; Benedict, 1973).

4.5.2 The Morphology of the Rock Glaciers in the Study Area

Figure 4.1 shows the location of the rock glaciers found within the study area. Tongue-shaped, ice-cored rock glaciers are found immediately downvalley from both of the cirque glaciers in the North and South Cirques. A spatulate rock glacier lies at the foot of Terminal Mountain (Plate 4.1), and lobate rock glaciers are found against the north flank of North Ridge. There is also an area below the ridge between Whistlers and Indian Passes which appears to be a rock glacier of considerable antiquity.

Most of the work in this study concentrated on the two tongue-shaped rock glaciers in the main cirques, and the spatulate rock glacier below Terminal Mountain. Figures 4.2, 4.3, and 4.4 are detailed plan views of each of these rock glaciers. The plan view of the South Cirque rock glacier (Fig. 4.2) also includes the ablation moraine described in Section 4.4.1. The figures illustrate the major

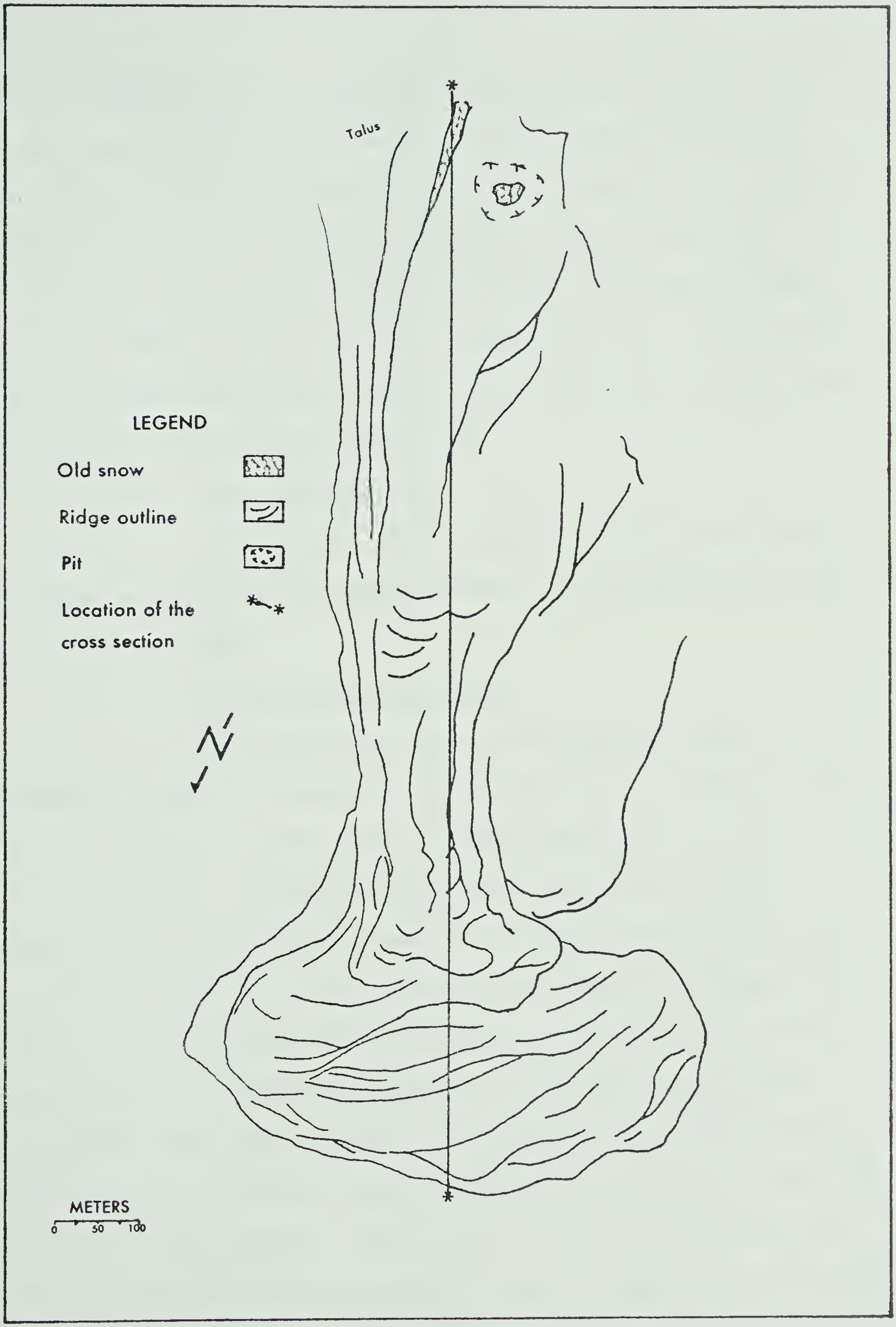


FIGURE 4.4 Plan view of the Spatulate rock glacier.

surface features and their respective lichen covers. However, owing to the large size and complex nature of the spatulate rock glacier, the lichen and weathering measurements were restricted to a 10 m wide transect running down the center of the rock glacier (Fig. 4.5c). Longitudinal profiles of all three rock glaciers are given in Figure 4.5a,b,c. Tables 4.1, 4.2, and 4.3 provide the details on the weathering values and lichen cover observed on the North Cirque, South Cirque, and spatulate rock glaciers,respectively.

A. Form and Microrelief

The rock glaciers in the study area all have slightly different morphologies and surface features. The following is a brief description of each.

North Cirque Rock Glacier

This rock glacier abuts the downvalley side of the prominent bedrock wall described in Section 4.3.2. It covers an area of ~0.2 km^2 and its overall form is tongue-shaped, with a peculiar assymetric front. The rock glacier is more extensive downvalley on the north side of the cirque where it is aligned with the gap in the bedrock wall (Fig. 4.2). The morphology and lichen data suggest that this rock glacier has two distinct lobes superimposed on one another. The lower lobe extends further downvalley than the overlying lobe. The profile (Fig. 4.5a) indicates the elevational differences between these lobes. The frontal slopes on both lobes are very steep (39^o-41^o), lichen free, and unstable. There is also a sharp break between the frontal slopes and the top surfaces of both lobes which lie at an angle

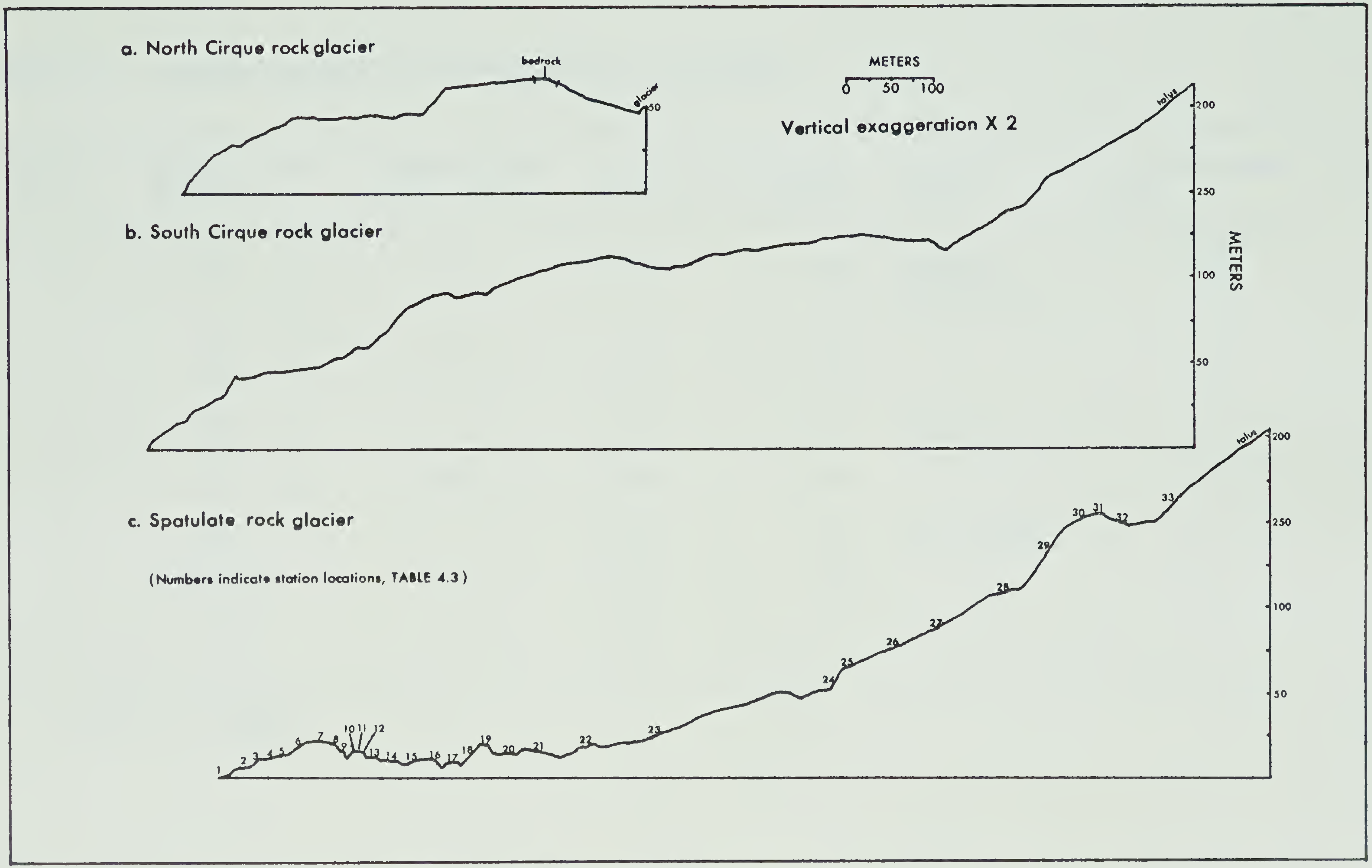


FIGURE 4.5 Longitudinal cross sections of the rock glaciers.

Table 4.1 Relative age data for the North Cirque rock glacier.

Location	Maximum Lichen Size (mm)	Percent Total Lichen Cover	Minimum Surface Age		Angularity of Debris	Rind Thickness (mm)
			A.D.	B.P.		
A	95	30	1375	575	angular-subangular	3
B	65	25	1700	250	"	2
C	72	30	1610	340	"	3
D	60	30	1770	180	angular	--
E	47	20 - 25	1850	100	"	--
F	20	1	1910	40	very angular	--
G	15(45*)	>1	1925	30	"	--
H	2	--	1945	5	"	--

*size of *X. elegans*

of ~10^{o}.

The top surfaces of these lobes are characterized by different microrelief. The microrelief on the lower lobe consists of longitudinal ridges, ~5 m in height, which curve into transverse orientations near the front. These successive ridges are roughly parallel to one another and the outermost one forms the lateral moraine described in Section 4.4.1 (Fig. 4.2). The slopes on some of these ridges are as steep and unstable as the frontal section of the lobes themselves, although they are much smaller. The microrelief on the upper lobe, on the other hand, consists of straight parallel ridges running the length of the lobes. These ridges are spaced ~30-60 cm apart and are separated by 'V' shaped furrows 20-30 cm deep. These features are composed of platy fragments of finely interlaminated shales and quartzites. The plates lying in the furrows are vertically aligned, whereas, on the intervening ridges, they are nearly horizontal. This alignment suggests that cryoturbation may have been responsible for their formation. However, if the nearby glacier overrode this surface, there is also the possibility that this surface became fluted by glacial action.

South Cirque Rock Glacier

The rock glacier in the South Cirque forms a continuum with the ablation moraine discussed in Section 4.4.1. The total area of this rock glacier-ablation moraine complex is ~0.4 km^2. As in the case of the North Cirque rock glacier this feature also has a tongue-shaped form with evidence of two lobes. However, in the South

Table 4.2 Relative age data for the South Cirque rock glacier. (cf. Fig. 4.3)

Location (Ridge)	Maximum Lichen Size (mm)	Percent Total Lichen Cover	Minimum Surface Age (A.D.)	Angularity of Debris	Rind Thickness (mm)
A	60	>50	1770 (180 BP)	angular	nil
B	35	10	1885	very angular	"
C	17	5	1920	"	"
D	37	30	1880	"	"
E	30	30	1898	"	"
F	28	5	1902	"	"
G	29(50*)	5-15	1909	"	"
H	30	15	1898	"	"
I	28	10-20	1902	"	"
J	35	5-10	1885	"	"
K	40	30	1875	"	"
L	20	10	1915	"	"
M	40	20	1875	"	"
N	35	3	1885	"	"
O	30	25	1897	"	"
P	55	50	1815	"	"
Q	50	50	1842	"	"
R	14	5-10	1925	"	"

*size of *X. elegans*

Cirque rock glacier the lower lobe has been largely overridden by the upper lobe and only a fragment of it is still visible (Fig. 4.3). The frontal slopes are again very steep (40^{o}) and unstable with a sharp break in slope as their upper surfaces are reached. The average slope of the entire rock glacier is $\sim 10^{o}$. Transverse, longitudinal, and lateral moraine-like ridges are found on the surface of this rock glacier. The longitudinal ridges are, however, much larger and wider than the ones on the North Cirque rock glacier, and continue on from the ridges described on the ablation moraine (Sec. 4.4.1, A).

A complex series of lateral moraine-like ridges are found along the west edge of the rock glacier (Fig. 4.3, ridges: J, K, L, M, N). These ridges are fragmented and the inner ones appear to truncate the outer ones. Further up the valley they become indistinguishable from the rock glacier-ablation moraine body. Another noteworthy feature is a large conical mound, rising ~5 m from the surrounding surface, near the front of the rock glacier (visible in Fig. 4.5b). The bulk of this mound is composed of a large block of finely interlaminated quartzites, slates, and shales. These lithologies are occasionally found interbedded with the massive Gog quartzites on the adjacent cirque walls. It is self evident that this block must have fallen from these cliffs at some time in the past and subsequently been transported to its present position. Additional surface features on the South Cirque rock glacier include sinuous furrows and conical pits. The sinuous furrows are especially prominent on the air photographs (Plate 4.1).

Spatulate Rock Glacier

Unlike the tongue-shaped rock glaciers which lie within well-developed cirques, the spatulate rock glacier is located at the base of an extensively subdued bedrock wall. No glacier occurs upslope from it nor is there any evidence of one in the past. Wahrhaftig and Cox (1959) have theorized that spatulate rock glaciers form when tongue-shaped rock glaciers flow onto a valley bottom. Consequently, the decrease in gradient causes a reduction in velocity which results in compressive flow and thickening and widening of the rock glacier. This theory may explain the widening of this particular rock glacier as widening does not occur until the relatively flat base of the valley is reached. The total area covered by this rock glacier is ~0.2 km^2. The microrelief on it consists of longitudinal and transverse ridges and furrows. The ridges are mainly longitudinal in the upper sections of the rock glacier but become predominantly transverse and arcuate as the rock glacier widens near its front. As this transition takes place,the ridges become contorted and the surface of the rock glacier becomes very irregular. These ridges (Fig. 4.5c) are much larger than the ones observed on the other tongue-shaped rock glaciers. There are no obvious lobes present on the surface of the spatulate rock glacier except for an area near the head, where the slope increases abruptly and becomes unstable (Fig. 4.5c) which may be an incipient lobe. Unlike the other rock glaciers described, the front slope of this rock glacier is stable, partially vegetated, and has an angle of ~10°. Along the west side of the spatulate rock

Table 4.3 Relative age data for the spatulate rock glacier.

Station	Maximum Lichen Size (mm)	Percent Total Lichen Cover	Minimum Surface Age		Angularity of Debris	Rind Thickness (mm)
			A.D.	B.P.		
1	50	--	1842	108	--	2
2	33	40-50	1890	60	angular-subrounded	2
3	20	25	1915	35	angular-subangular	1
4	75 (240*)	> 50	1585	365	"	2
5	--	--	--	--	--	--
6	40	> 50	1875	75	angular	2
7	140	> 50	1015	935	very angular-subangular	5
8	90	> 50	1455	495	"	1
9	100	> 50	1370	580	"	1
10	45	50	1860	90	"	1
11	80	> 50	1545	405	angular	1
12	90	> 50	1455	495	--	0
13	135	> 50	1060	890	angular	1
14	120	> 50	1195	760	angular-subangular	2
15	30	25	1898	52	very angular subangular	0

Table 4.3 Continued.

Station	Maximum Lichen Size (mm)	Percent Total Lichen Cover	Minimum Surface Age		Angularity of Debris	Rind Thickness (mm)
			A.D.	B.P.		
16	120	> 50	1195	760	angular-subangular	2
17	125	50	1150	800	"	0
18	--	--	--	--	very angular-angular	--
19	85	> 50	1500	450	angular	0
20	50	20	1842	108	very angular-angular	0
21	60	> 50	1770	180	angular-subangular	0
22	70	> 50	1630	320	angular	2
23	110	> 50	1280	670	angular	1
24	--	--	--	--	--	--
25	60	40-50	1770	180	angular	0
26	31	25-30	1897	53	angular	0
27	25	20	1907	43	very angular-angular	0
28	25	3	1907	43	"	0
29	--	--	--	--	--	--

Table 4.3 Continued.

Station	Maximum Lichen Size (mm)	Percent Total Lichen Cover	Minimum Surface Age		Angularity of Debris	Rind Thickness (mm)
			A.D.	B.P.		
30	60	50	1770	180	angular	0
31	65	30-40	1700	250	very angular-angular	--
32	--	--	--	--	--	--
33	35	--	1885	65	very angular-angular	0

*size of *X. elegans*

glacier lies a small mass of coarse rubble with faint transverse ridges on its downslope side. This feature is thought to be a rock glacier of very minor development.

Other Rock Glaciers

The remaining rock glaciers in the field area were studied only in brief detail. Several coalescent, lobate rock glaciers extend along the entire north side of North Ridge (Fig. 4.1). The easternmost rock glacier was found to have a similar lichen cover and composition as the spatulate rock glacier below Terminal Mountain. The front slope of the North Ridge rock glacier also lies at 10°, has a turf cover, and does not exhibit multiple lobes.

A second area briefly studied occurs below the ridge which separates Indian and Whistlers Passes (Fig. 4.1). This area is characterized by coarse debris covered with turf and shrubs. The terrain is hummocky and has the appearance of transverse ridges such as those found on the other rock glaciers in the study area. For this reason it is speculated that this may be a rock glacier complex, but one of much greater age than any of the other examples found in the valley. Alternatively, it may be the remnants of the drift sheet described in Section 4.4.2, or the product of a large rock slide from the ridge above, hence its identification remains tentative.

B. Composition

Rock glaciers must have a certain amount of ice to maintain movement. As noted in Section 4.5.1, this ice may be in the form of a buried ice-core or interstitial fillings. Also noted

previously, buried glacier ice was observed in the ablation moraine occupying the South Cirque (Sec. 4.4.1, A). It is reasonable to assume, therefore, that this ice continues downvalley and lies buried in the rock glacier sector, under a thicker cover of debris. Hence, it is probable that this rock glacier is of the ice-cored type. No ice was observed in the North Cirque rock glacier. However, because it appears to be active and also occurs immediately downvalley from a contemporary glacier, it is assumed that it contains an ice-core. The spatulate and lobate rock glaciers, on the other hand, show no sign of movement so the amount of ice within them, if any, must be minimal. The differences in ice content between the spatulate rock glacier and the rock glaciers occupying the cirques may be reflected by their relative longitudinal profiles. The active rock glaciers in the cirques have a convex profile, whereas the stabilized spatulate rock glacier has a concave profile, suggesting a prior volume of now melted out ice. The relative amounts of ice within these rock glaciers may also be indicated by the amounts of water issuing from their fronts. The active rock glaciers have substantial discharges, especially the South Cirque rock glacier where a large stream exits from its base. The spatulate rock glacier, on the other hand, has only a few small springs. In addition, the spatulate rock glacier also differs from the other rock glaciers in that there is no indication of a past ice glacier at its head. This suggests that a different process was responsible for its formation, perhaps by interstitial ice.

All the rock glaciers in the study area are composed

of similar lithologies, with their upper portions lying at the base of Gog quartzite cliffs. On the other hand, rock glaciers are notably absent from slopes composed of the Upper Miette Formation (argillites, siltstones, sandstones). As noted in Chapter II, the Gog quartzites are massive and break into large equidimensional blocks up to 4 m in size which may be a factor in preserving the ice. The debris on the active rock glaciers in the study area tends to have a two layered structure: a coarse upper layer lacking in fines, and a lower layer composed of smaller clasts mixed with fines (Plate 4.6). The fines have a loamy sand to sandy texture (Appendix B) and are visible under a few meters of coarse material along the frontal slopes of the rock glaciers. Potter (1972) has suggested that such crude sorting may be due to: 1) windblown fines sifting down through the coarse surface rubble; 2) meltwater carrying the fines down to the lower layers; or 3) the *in situ* abrasion of the lower blocks, producing fines as movement takes place.

Another type of sorting is found in a number of the surface ridges on the rock glaciers, especially the spatulate rock glacier, where some ridges are largely composed of uniformly-sized debris. This may be due to frost sorting, however, it seems more likely to be related to the nature of the debris cones which feed some of these ridges. Some talus cones on the headwalls produce relatively uniform debris which is probably the result of local variations in bedrock structure (joints). The best examples of this are associated with the finely interlaminated slates and quartzites found in scattered

Plate 4.6 A side view of the South Cirque rock glacier (North is toward the upper left hand corner). The two-layered structure of the rock glacier is clearly discernable: a few meters of coarse, surface rocks, underlain by finer material. Also note the tree covered terrain in contact with the front of the rock glacier, two lobes at the front of the rock glacier, and a conical mound on its top surface (top center of the photograph).

outcroppings along the headwalls. The talus cones produced by these rock units consist of platy fragments only a few centimeters in length, and are dark in tone. Hence, these contrast sharply with the light colored quartzite blocks making up the surrounding debris.

C. Evidence of Movement

Wahrhaftig and Cox's (1959) criteria for determining whether a rock glacier is active are based on the characteristics of its front slope. The criteria for active rock glaciers are as follows: 1) the frontal slope of the rock glacier is unvegetated; 2) the slope lies at the angle of repose; 3) it has exposed fines; and 4) there is a sharp break between the frontal slope and the top surface of the rock glacier. Although these criteria are generally true, recent evidence has shown that there are exceptions (Luckman and Crockett, 1978). Whalley (1974) has also suggested that a similar morphology might be maintained if a debris mantle keeps resettling as a buried ice-core melts out, and therefore, no forward movement is necessary. Wahrhaftig and Cox (1959) reasoned that in order to maintain a sharp angle between the surface and front slope of a rock glacier, the debris on the top surface must be moving faster than the debris underlying it. In addition, there must also be a constant resupply of debris to the front of the rock glacier otherwise the steep frontal slope would diminish and the sharp break in slope would round off through mass wasting. They also concluded that in order to keep fines exposed along the front slope, the rest of the rock glacier must also be advancing internally. Based on Wahrhaftig and Cox's

(1959) criteria, the rock glaciers in the North and South Cirques are active whereas the spatulate and lobate rock glaciers are inactive.

The best evidence for rock glacier advancement in Whistlers Creek Valley occurs along the front of the rock glacier flowing out of the South Cirque,where vegetation is being overridden (Plates 4.7 and 4.8). In this location, trees growing adjacent to the rock glacier are tilted away from it at various angles, and the bases of some of these trees have been recently buried. Other trees, up to 3 m away from the front, are freshly scarred and broken by large blocks which have rolled off the top of the rock glacier. Overturning some of these stones has revealed recently buried heaths and mosses, and in a number of cases, the underside of the blocks are lichen covered. Also in some cases, the rocks at the base of the rock glacier appear to be ploughing and overturning turf before it. All the above evidence suggests very recent activity.

As noted in the previous chapter, the initial tilting of a tree can be determined by an examination of its tree rings (i.e., the formation of reaction wood). A number of the tilting trees in the front of the South Cirque rock glacier were cored. The results of the coring are given in Table 4.4. These trees were found either partially buried at the base of the rock glacier or in a tilted attitude up to 2 m away from its front. Evidence from seven trees cored ~1 m from the front (Table 4.4, T-6 to 12) indicates that they began tilting from 13 to 31 years ago, even though they are not in physical contact with the rock glacier. This tilting is thought to be induced by the

Plate 4.7 The steep front of the South Cirque rock glacier.

Plate 4.8 Tilted trees and overridden vegetation at the front of the South Cirque rock glacier.

Table 4.4 Tree core data. (for location see Fig. 3.2)

Sample no.	Species	Age	Distance from rock glacier front	Indicated time of tilt (years ago)	Comments
T-1	Alpine fir	133	- 1 m	45	Buried by the rock glacier.
T-2	"	108	- 50 cm	40	"
T-3	"	143	0	--	No reaction wood found.
T-4	"	44	75 cm	13	
T-5	"	54	1.5 - 1 m	19	
T-6	"	95	"	25	Tree broken and dead.
T-7	"	89	"	22	
T-8	"	36	"	17	
T-9	"	107	"	21	Badly bruised tree.
T-10	"	102	"	57	Reaction wood only for 9 years.
T-11	"	88	"	13	
T-12	"	166	"	31	
T-13	Spruce	288	--	--	Growing in the forest before the South Cirque rock glacier.
T-14	"	322	--	--	"
T-15	"	243	--	--	"
T-16	Alpine fir	44	--	--	Growing on the spatulate rock glacier.

loading effects of snowbanks which accumulate against the front of the rock glacier during the winter. One particular tree studied (T-1) was found with its base buried 1 m behind the front of the rock glacier. The tree ring data obtained indicates that it began tilting about 45 years ago. If this particular tree began tilting when the front of the rock glacier was from 1 to 2 m away, then an approximate rate of advance can be calculated (2.2-4.4 $cmyr^{-1}$). This is, of course, only a minimum estimate since the movement at the front of this rock glacier is likely to be sporadic as stones roll off the front face (i.e., see sporadic movement of the Arapaho rock glacier, Table 4.5) and it is also possible that initial tilting did not occur when the tree was 1-2 m away from the front edge. A comparison of this local rate of advance to other rates described in the literature (Table 4.5) suggests that this value is slower than average, perhaps because this rock glacier must flow slightly uphill as it advances onto the vegetated surface.

Since the front of the rock glacier in the North Cirque lies above timber-line, no similar evidence exists to demonstrate that this rock glacier is advancing. However, inspection of stones in front of this rock glacier has shown that a number of them have been recently deposited, as indicated by lichens on their undersides,and the absence of lichens on their presently exposed surfaces. In addition, the upper lobe of this rock glacier has a number of features which look like slump scarps near the front of it which may indicate recent settlement as the ice-core is melting out (Fig. 4.2).

Table 4.5 Some recorded rates of movement for rock glaciers.

Source	Location	Rate ($cmyr^{-1}$)
Wahrhaftig and Cox (1959)	Clear Creek Rock Glacier, Alaska	62
Outcalt and Benedict (1965)	Arapaho Rock Glacier, Colorado	19
White (1971)	Arapaho Rock Glacier, Colorado;	5
	Taylor Rock Glacier, Colorado;	6.6
	Fair Rock Glacier, Colorado	9.7
Osborn (1975)	Lake Louise, Banff National Park	70-80
Hughes, Rampton, and Rutter (1972)	Sheep Mtn. Rock Glacier, Yukon Territory	35
This Study	South Cirque Rock Glacier, Whistlers Creek	>2.2-4.4

Rock glacier activity may also be indicated by the general lichen development found on their surfaces. On the rock glaciers studied it was found that lichens progressively decrease in size and coverage as one goes from the terminus, to the head; or from the sides, toward the center (Fig. 4.2, 4.3, Tables 4.1, 4.2). The lichens are also usually restricted to the ridge crests and are commonly absent in the furrows. If lichens colonize a surface as the substrate becomes stable, two alternative interpretations are warranted. First, it appears that stabilization of the debris has taken place, initially beginning at the front and sides and working progressively towards the center. This situation would suggest that there is a stagnant ice glacier buried under the debris which is melting out from the edges towards the center. The second interpretation is that the lichen distribution represents the amount of time the surface has been exposed as it travelled from the headwall source. The largest lichens, therefore, would be found at the front of the rock glacier since this is where the greatest distance to the headwall occurs. The best support for this interpretation is found along the longitudinal ridges which grade into talus cones. Here lichens are absent near the cliffs, but progressively increase downvalley along the ridge. The lichen distribution found on the rock glaciers in Whistlers Creek Valley is probably the result of both of the above factors. In addition, certain areas on the surface of the rock glaciers have a thicker debris mantle, and hence greater stability, so lichens colonize these areas more quickly, whereas other areas with less surface debris

are subject to instability as the ice melts out (i.e., the pits and furrows). A detailed discussion of the past movement of the rock glaciers in the study area is included in the next chapter.

4.5.3 Summary and Conclusions

The study area contains rock glaciers of several morphological types. Each type has specific characteristics which suggest different modes of formation. The tongue-shaped, ice-cored rock glaciers are found in cirques, whereas the spatulate and lobate, ice-cemented rock glaciers are found at the base of steep cliffs. The evidence presented for the tongue-shaped rock glaciers in the North and South Cirques suggests that the ice glaciers immediately upvalley were instrumental in their formation. Their development was likely related to an initial advance of the ice glaciers during which the glaciers were capable of transporting the high input of detritus from surrounding cirque walls. However, as the climate ameliorated and/or aridity increased, the glaciers retreated and were no longer capable of transporting the material falling onto their surfaces. This resulted in an ever increasing debris cover on the glaciers which would have been augmented by the accumulation of englacial debris. As the glaciers thinned, the supply of ice to the front was terminated and the ice became stagnant. Stagnation of the glacier fronts was probably greatly enhanced by the bedrock obstructions occurring in both of the cirques which effectively cut-off the source areas. Eventually the debris mantle on these buried glaciers became sufficiently thick that it insulated the stagnant ice from completely melting.

Under the accumulating overburden, the remnant ice-cores were able to flow and advance whereas the ice glaciers immediately above them retreated.

On the other hand, the spatulate rock glacier at the base of Terminal Mountain probably had a different origin. It is thought that this rock glacier was formed when the climate was severe enough for ice to accumulate in the interstices of coarse talus at the base of the mountain. These north-facing cliffs presently offer maximum protection from solar radiation and may have developed permanent snow patches in the past. Such snow patches would provide moisture to aid interstitial ice formation, and in turn, may also have been directly buried by debris and subsequently incorporated into the rock glacier. Eventually the ice content became such that the ice-rock mixture flowed down the valley to its present position.

At present, the ice-cored rock glaciers are active whereas the spatulate rock glacier is inactive. This may suggest that the initial climatic deterioration which eventually produced the ice-cored rock glaciers was not severe enough to initiate the development of the ice-cemented type. A similar observation was made by Luckman and Crockett (1978) for a number of sites within Jasper National Park. In addition, the head of the spatulate rock glacier is ~180 m lower than the heads of the active rock glaciers which may indicate that the earlier phase of rock glacier development was climatically more severe than the most recent one (because of the lower paleoshowline implied).

Lobes form on rock glaciers when they become rejuvenated

after a period of inactivity. Two distinct lobes were found on both of the ice-cored rock glaciers and an incipient lobe was found on the spatulate rock glacier. It is possible that the lobes on the ice-cored rock glaciers formed when their adjacent glaciers advanced and overrode them, only to stagnate again as described above. This mechanism may also explain the complex series of lateral moraine-like ridges found in a few sections of the rock glaciers. Development of a lobe on the spatulate rock glacier may have been curtailed during the latest climatic deterioration since the earlier phase of rock glacier activity appears to have denuded the cliffs to such an extent that the supply of detritus to the head has been exhausted. Complex forms of microrelief found on the surfaces of the rock glaciers are believed to relate to internal flow and to differential melting of ice rich areas, enhanced by heat conduction from meltwater.

4.6 Slope Debris Deposits

A very wide range of landforms fall into this category, including the rock glaciers discussed in the previous section. This category is divided into three basic sections: avalanche and talus slopes; solifluction landforms and patterned ground; and ploughing blocks.

4.6.1 Avalanche and Talus Slopes

Most of the debris at the base of the quartzite cliffs originates from either rockfalls or avalanches (Fig. 4.1), forming aprons and cones. As noted in the previous section, these features often feed the heads of the rock glaciers. The cones and aprons which are composed of a wide range of rock sizes, including interstitial

fines, are thought to have a polygenetic origin, whereas those composed only of coarse debris are thought to result primarily from rock falls. Scars from former mudflows, often with well developed levees on their banks, are common on these deposits. Protalus mounds of recent origin, formed by the accumulation of debris at the base of snowbanks, were also observed in the upper cirques. A number of block fields were also found on the valley floors (Fig. 4.1).

Based on lichen cover existing on some of the above deposits (Table 4.6) it is evident that a number of these slopes are being constantly renewed, whereas others are inactive. Formation of these have resulted from either a more severe climate, with increased freeze-thaw activity, or simply from debris provided by cliffs which have become exhausted. Morphological evidence suggests that both conditions exist in the study area.

4.6.2 Solifluction and Patterned Ground

Solifluction (gelifluction) is the most ubiquitous process in the area and a wide variety of associated landforms exist. The most common are lobes and terraces (Fig. 4.1).

One of the most interesting locations for solifluction lobes occurs at the snouts of the spatulate and lobate rock glaciers previously discussed. The lobes at the base of the spatulate rock glacier (Fig. 4.1) are of the turf and stone-banked type (cf. French, 1976). The stone-banked lobes have turf on their top surfaces at an angle of 17-18^{o}, whereas the risers are commonly at 55^{o}. One particular stone-banked lobe was excavated and examined. The

Table 4.6 Relative age data for selected landforms and surfaces in Whistlers Creek Valley.

Site	Maximum Lichen Size (mm)	Percent Total Lichen Cover (mm)	Minimum Surface Age A.D.	Minimum Surface Age B.P.	Angularity of Debris	Rind Thickness (mm)
1. Rock glacier form (adjacent to the spatulate rock glacier)	120	50	1195	760	angular-subangular	5
2. Lobate rock glacier (North Ridge)	120	50	1195	760	angular-subangular	2
3. Lateral moraine (North Cirque)	130	> 50	1105	845	"	3
4. Lateral moraine (North Cirque)	38	5	1880	70	very angular	0
5. Lateral moraine (North Cirque)	0	0	< 1950	--	"	0
6. Terminal moraine (North Cirque)	65	> 50	1700	250	angular-subangular	3
7. Terminal moraine (North Cirque)	65	> 50	1700	250	"	3
8. Bedrock surface (adjacent 3)	--	--	--	--		3
9. Bedrock surface (adjacent 4)	50	> 50	1842	108		1

Table 4.6 Continued.

	Maximum Lichen Size (mm)	Percent total Lichen cover (mm)	Minimum Surface Age A.D.	Minimum Surface Age B.P.	Angularity of Debris	Rind Thickness (mm)
10. Bedrock surface (in front South Cirque rock glacier)	55	> 50	1815	135	--	2
11. Bedrock surface (between Indian and Whistlers Passes)	120	> 50	1195	760	--	5
12. Tor (Marmot Mtn.)	130	--	1105	845	subangular-subrounded	6
13. Talus (North Cirque)	33	40	1890	60	angular-subangular	0
14. Talus (North Ridge)	65	> 50	1700	250	"	7
15. Talus (Whistlers Pass)	75	--	1585	365	subangular	7
16. Talus (North Ridge)	--	> 50	--	--	subangular	4
17. Blockfield	--	50	--	--	angular-subangular	5

riser portion of this lobe, ~1 m high, was composed of large blocks of various sizes stacked on each other, with an absence of fines. Orientation of the clasts was not evident because of their equidimensional nature. No signs of recent movement such as recently buried vegetation at the base of the lobe was evident. However, the top parts of some of the lobes were disturbed by recent mudflows of a localized nature which washed fines between the rocks, mobilizing them and rupturing the turf. The turf-banked lobes examined were commonly associated with springs which produce saturation. Excavation of one particular lobe showed a surface layer of turf overlying fine silts. Below 12 cm a heterogeneous mixture of coarse rock debris, including fines, was encountered. Recently overridden vegetation under the front of the lobe suggests that it is active. No organic material, however, was collected for radiometric dating as the material was apparently of such recent origin.

One large, turf-banked solifluction lobe was found at the front of the lobate rock glacier (Fig. 4.1). The turf covered tread of this lobe, lying at ~15^{o}, is composed of a mixture of fines and rock clasts which contrast markedly with the large blocks composing the rock glacier immediately behind it. On the front surface of this lobe there are a number of places where the continuous turf cover has been recently ruptured. This may suggest a recent increase in solifluction processes acting on the lobe. The similarity in location of the above lobes suggests that they formed as fines accumulated at the front of these inactive rock glaciers. These fines likely accumulated as springs and meltwater washed through the rubble of the upper rock

glaciers and exited at their fronts. A number of springs are presently emerging from the front of the spatulate rock glacier and these provide abundant moisture for solifluction.

Other solifluction lobes and terraces were found on slopes ranging from 8^{o} on Marmot Pass to 34^{o} on the northwest flank of Marmot Mountain. Solifluction terraces were especially prominent on the soggy southeast slopes of Indian Ridge (11^{o} slope, Fig. 4.1). The location of solifluction in the study area did not seem to depend on slope orientation, but rather on the abundance of moisture and a substrate of fines.

Patterned ground in the study area consisted of sorted and non-sorted stripes, sorted and non-sorted polygons, sorted steps, turf hummocks, and turf rimmed nets. With the exception of the turf hummocks found in the North Cirque, all the patterned ground was associated with the small platy fragments of the Upper Miette Formation (for location see Fig. 4.1). All these forms were fresh in appearance and in some cases showed signs of very recent activity. No relic patterned ground was found. Excavation of the turf hummocks in the North Cirque (Fig. 4.1) indicated frozen ground 20 cm below the surface by mid-August of the 1977 field season, and may be evidence of permafrost at this elevation (~2350 m).

4.6.3 Ploughing Blocks

Ploughing blocks, sometimes known as gliding blocks, are a product of solifluction, particularly frost creep. They are boulder-sized stones which move down a slope leaving a furrow behind them, and

forcing up ground before them (Plate 4.9). Tufnell (1972, 1976) has extensively studied the movement of ploughing blocks in the Moor House National Reserve in England. Tufnell found that 95% of the movement occurred during the winter and total movement was commonly <8 $cmyr^{-1}$. Block movement varied from year to year depending on both the severity of the winter (a severe winter would accelerate movement), and on changes in slope gradient encountered by the block as it moved downhill. Tufnell concluded that ploughing blocks were a product of a mild periglacial environment as their movement was predominantly produced by frost action (heaving) assisted by gravity, water, and temperature fluctuations $>0^{o}C$, causing expansion and contraction of the blocks which result in a net downhill creep.

A number of ploughing blocks were found in the study area (Fig. 4.1). Since ploughing blocks must move faster than the surrounding slope material, they are commonly found on stable, vegetated terrain. A few ploughing blocks in the study area, however, have also been found on non-vegetated slopes. Some non-vegetated surfaces also contain "brake blocks" (Tufnell, 1972). In this case the surrounding fines move faster than the block so that a furrow forms on the downhill side.

One of the most interesting locations where ploughing blocks are found is within a forested area northeast of the North Ridge (Fig. 4.1, ~10^{o} slope). The blocks found in this area are up to 4 m in length, the largest of which had a pronounced mound ~75 cm high on its downhill side and a deep furrow ~ 20 m long on the uphill side.

Plate 4.9 Ploughing blocks. Note the pushed up earth at the front of the block and the trailing furrow.

The furrow was completely covered with turf, except for a 10-20 cm gap of non-vegetated silt immediately behind the block. In addition, tree and shrub vegetation is noticeably absent in the furrow,whereas shrubs are well established along both sides of it.

Several conclusions may be made from the observations on this latter block. First, the length of the furrow behind the block gives a minimum estimate (~20 m) of the total movement that the block has travelled. Secondly, insofar as the furrow consists of a vegetated and non-vegetated surface, several phases of movement must have taken place. The most recent movement (non-vegetated silt zone) was at least 10-20 cm and may have occurred during the previous winter. Thirdly, it is possible that this block has overridden dateable organic material which now remains buried in the furrow.

Another interesting example of ploughing blocks was found on Marmot Mountain. Here it appears that movement was initiated as a solifluction lobe advanced beneath a block and increased its gradient. The entire length of the furrow (~1 m) was non-vegetated suggesting one, single, recent movement.

In summary, a large number of ploughing blocks have been found in the study area. These blocks are commonly found on gentle slopes (~10^{o}) and show evidence of multiple movements. Some of the large blocks are ploughing through vegetated terrain and may have dateable organic material buried in the furrows behind them. Although some work on ploughing block movements has been done in Europe (i.e., Tufnell, 1972, 1976), the work in North America has been virtually

non-existent. However, research on the past movements of ploughing blocks, such as those described above, may reveal periods of increased periglacial activity that occurred in the past.

CHAPTER FIVE

LIMITATIONS AND RESULTS OF THE RELATIVE AGE STUDIES

In this chapter the limitations and results of the relative age dating techniques described in Chapter III are discussed. In addition, the recent glacial and periglacial history of Whistlers Creek Valley is tentatively outlined from the evidence provided.

5.1 Lichenometry

The lichenometric data for various surfaces and features in the study area are found in Tables 4.1, 4.2, 4.3, and 4.6. Lichenometry proved to be a very useful technique for interpreting and dating recent movements in the rock glaciers. It was also found that certain lichen species such as *Umbilicaria deusta*, *Acarospora glaucocarpa*, and *Alectoria pubesens* dominate the lichen cover in the study area to the extent that the quartzitic cliffs and debris slopes are literally blackened by them. Since the substrate is light colored, distinctions between very fresh surfaces and older, lichen covered surfaces can easily be made. However, there were a number of limitations when attempting to date older surfaces. One such problem is that where the black lichen *Umbilicaria deusta* and *Acarospora glaucocarpa* are found with *R. geographicum*,it appears that they are able to dominate and grow over the Rhizocarpon. It was observed that surfaces which were considered "old" according to other relative age criteria, including a very extensive lichen cover, had relatively small *R. geographicum* thalli. For example, the moraines below the North Cirque rock glacier

(Fig. 4.2) have maximum thalli of 65 mm, whereas the rock glacier itself has a maximum of 95 mm (Table 4.6). The dominance of the other lichens then, probably hinders optimum growth of *R. geographicum*, especially as the substrate becomes increasingly crowded through time. This, in turn, leads to increasing discrepancies between actual age and the age determined by the lichen growth curve.

Figure 5.1 is a graph of the relationship between maximum *R. geographicum* thalli and the percent of total lichen cover found on various surfaces in the study area. In general, the graph shows that a surface is quickly covered by lichens as the size of *R. geographicum* increases. Approximately 50% of the surface is covered by the time the thalli reach 65 mm. Omitting extremes, and values of >50% cover (the limit of measurable cover by the technique described in Chapter III), a linear regression indicates that the theoretical maximum diameter of *R. geographicum* is 125 mm (as this is when 100% lichen cover is achieved). However, this relation is unrealistic because the lichen thalli are roughly circular and contact between adjacent thalli would actually occur long before 100% lichen cover has been reached. In addition, the actual rate at which a surface becomes covered with lichens will vary depending on the number of lichens and the types of species growing on that surface, as well as on the microclimatic conditions. Typically, measurements in the field have shown that after ~ 50 mm of growth,interference between lichens occurs and individual *R. geographicum* thalli tend to coalesce. The above evidence then, suggests that using very large, isolated thalli for dating may

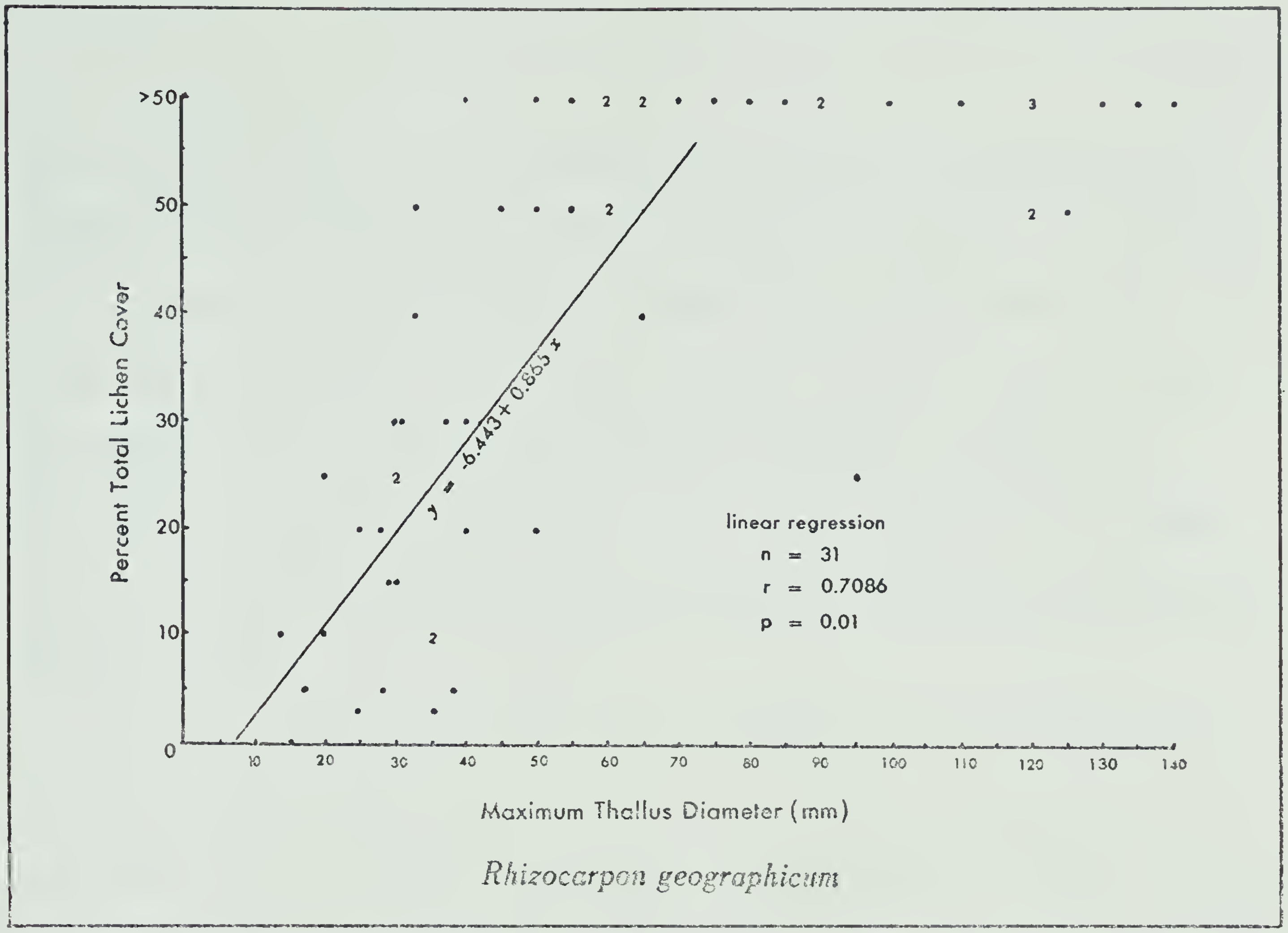


FIGURE 5.1 Graph of maximum lichen thallus versus percent total lichen cover.

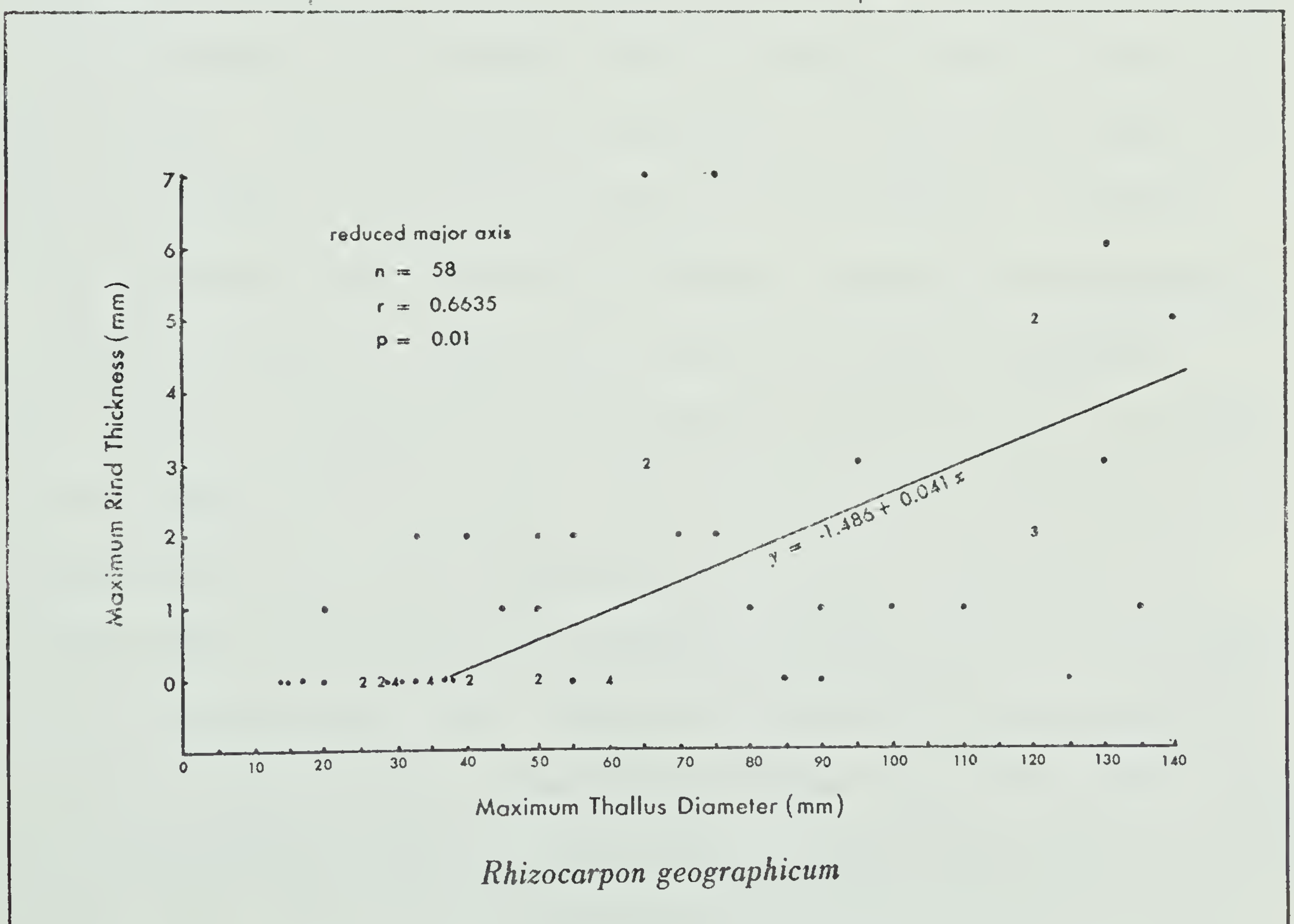


FIGURE 5.2 Graph of maximum thallus versus weathering rind thickness.

be erroneous. The very large thalli are probably not growing in the optimal microenvironments necessary for dating (see Sec. 3.2.3) because if they were, interference with adjacent lichens would have long hindered their growth. In addition, Figure 5.1 also shows that the percentage of lichen cover increases very rapidly in relation to increasing *R. geographicum* thalli. It is, therefore, of less value than *R. geographicum* size in making relative age distinctions between older surfaces.

Another problem that occurs when attempting to date the older surfaces with *R. geographicum* involves the possible senescence of the species. It is not known if, or when, *R. geographicum* reaches a maximum size before growth ceases and deterioration occurs. Many of the larger lichens measured were partially eroded, especially near the center of the thallus. This may indicate senescence or, on the other hand, it may just reflect the harsh environment of the study area. Blowing snow and debris, due to high wind velocities coupled with depauperate vegetation, result in substantial abrasion in alpine environments. The largest *R. geographicum* thallus found in the study area was 140 mm, on the spatulate rock glacier. Older surfaces (such as the tors and other bedrock areas) had maximum sizes of 130 mm. Therefore, 130-140 mm thalli may represent the maximum attainable size of *R. geographicum* in the study area. The limiting factors may be the ones discussed above or, alternatively, it may be that at some time in the past, mechanical weathering may have been severe enough to have destroyed all previous lichen cover. Hence, the lichen sizes would represent the time since

this accelerated weathering had ceased. However, Luckman (1977) found *R. geographicum* thalli up to 280 mm in the Mt. Edith Cavell area and workers in other areas have also found larger specimens (i.e., Benedict, 1967; Denton and Karlen, 1973).

A third major limitation for dating the older surfaces in the Whistlers Creek area arises from the fact that the growth curve for *R. geographicum* (Luckman, 1977, Fig. 3.1) extends back only 250 years. Luckman (1977) suggested that the growth rate beyond 250 years is ~11.4 mm per 100 years, however,no data exists so age estimates must remain speculative.

As an aside, the *Xanthoria elegans* specimens found in the study area were dated using Osborn and Taylor's (1975) growth curve for the Lake Louise area. When compared with the dates obtained from *R. geographicum* for the same surfaces, wide discrepancies occurred. These disagreements probably emphasize the effects of differing environments between Lake Louise and Whistlers Creek.

In summary, when dating surfaces by lichenometry the most recent age estimates, back to ~250 BP, are thought to be reasonably accurate. Also, obvious distinctions could be made between very recent and older deposits based on percent lichen cover. Despite the problems discussed above, the lichen data for the study area is internally consistent. That is, with few exceptions, the size relationships of *R. geographicum* agreed with the relative ages of the surfaces, as indicated by cross-cutting relationships and superposition. Nonetheless, lichenometric dates beyond 250 BP must be regarded as very rough approximations.

5.2 Rock Weathering

The rock weathering data is presented in Tables 4.1, 4.2, 4.3 and 4.6. Oxidation rinds were the only weathering feature which could be systematically sampled in the study area, although minor differences in angularity of the rocks were detected on debris of obviously differing age. Weathering pits and crystal relief were almost absent on the deposits observed, except on some conglomerates (i.e., the tors). This was largely due to the resistant nature of the quartzite rocks in the study area and it is apparent that considerable time is required for any noticeable weathering differences to occur (cf. Dyke, 1978). Figure 5.2 is a graph showing the relationship between rind thickness and lichen diameter (*R. geographicum*) found in the study area. A significant positive correlation is indicated ($r = .66$, $p = .01$) between the two variables. Since there is no clearly defined dependent variable, a reduced major axis line was calculated to describe this relationship (Till, 1974). The line indicates that no detectable rind forms until the lichen are ~40 mm in size. This corresponds to ~70 years, based on the lichen growth curve. Figure 5.2 also shows that there is generally a maximum rind depth found for any given *R. geographicum* size. However, it is possible to have a number of thinner weathering rinds for the same sized lichen in other locations. In some cases heavily lichen covered areas had no rind at all. A number of microenvironmental conditions, including rock type and constituent grain size, govern the extent of rind development (cf. Birkeland, 1973). Some lichens were found to be leaching the surface of the rock, producing light colored

rinds which might accelerate the formation of oxidation rinds. The two extreme values of 7 mm were found on fairly coarse grained conglomerates. In addition, mechanical weathering such as exfoliation may counter the preservation of rinds through time and give erroneous age estimates. The above problems were limited as much as possible by sampling enough stones at a given site until what was considered to be a representative sample was obtained. Although the limitations described hindered differentiation of subtle changes in rock weathering, gross distinctions of weathering zones could be made, especially going well beyond the range of lichenometry.

5.3 Establishing a Chronology

This section further interprets the lichen and weathering rind data discussed in the previous section and establishes a chronology based on this data. Figure 5.3 shows a breakdown of the lichen data (found in Tables 4.1, 4.2, 4.3 and 4.6) into a histogram. When considering just maximum *R. geographicum* values for selected sites in the study area, there appears to be a trimodal distribution of lichen sizes. The modes are 31-40 mm, 51-70 mm, and 111-130 mm, respectively. When additional data of maximum thalli are added, such as those found on the ridges of the rock glaciers, the modality appears roughly the same (Fig. 5.3). This distribution suggests at least three different periods of increased lichen colonization. The two youngest phases fall within the range of relatively accurate lichenometric dating. The most recent phase dates in the late 1880's, whereas the preceding

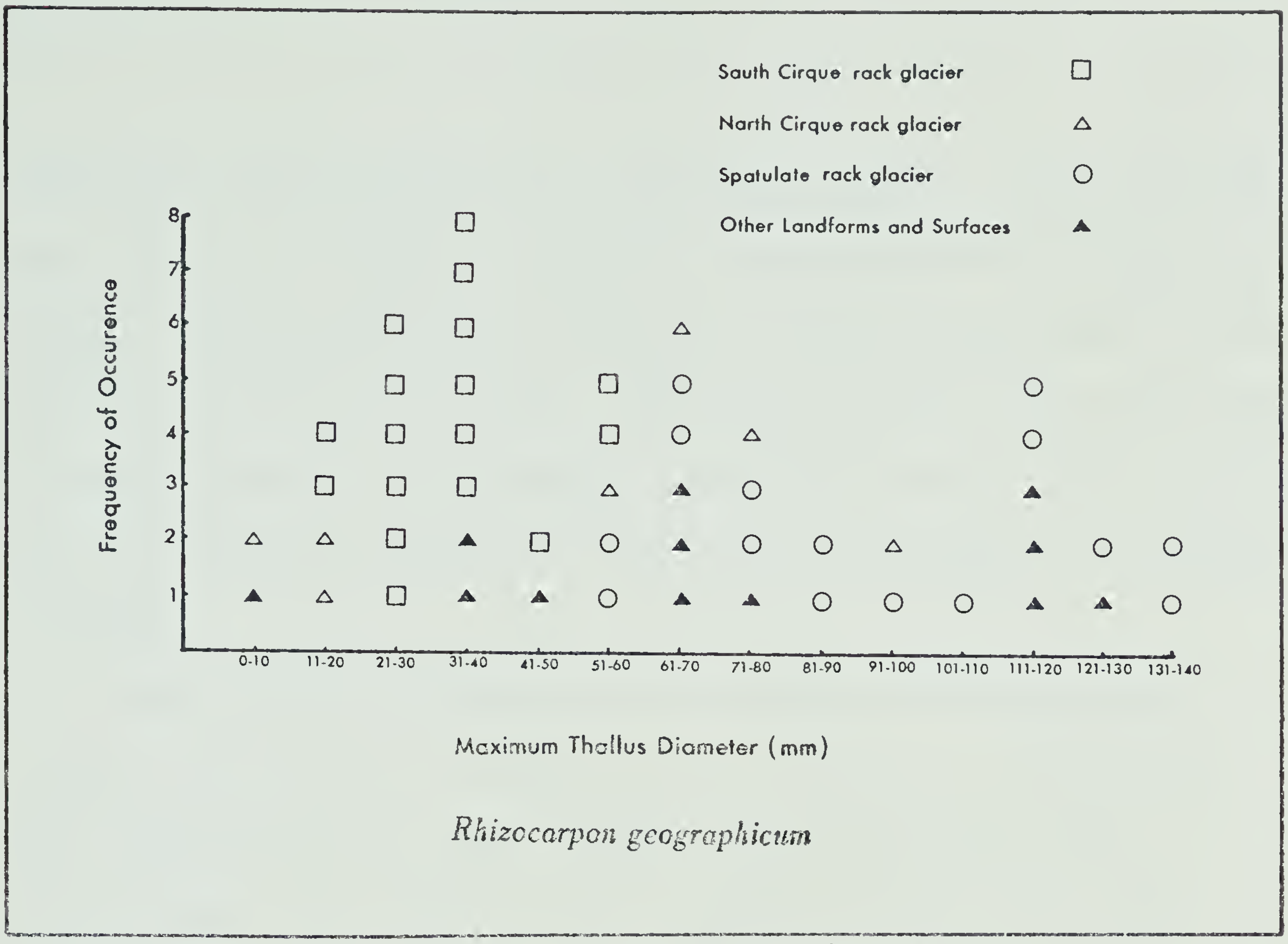


FIGURE 5.3 Histogram of lichen thalli in the study area.

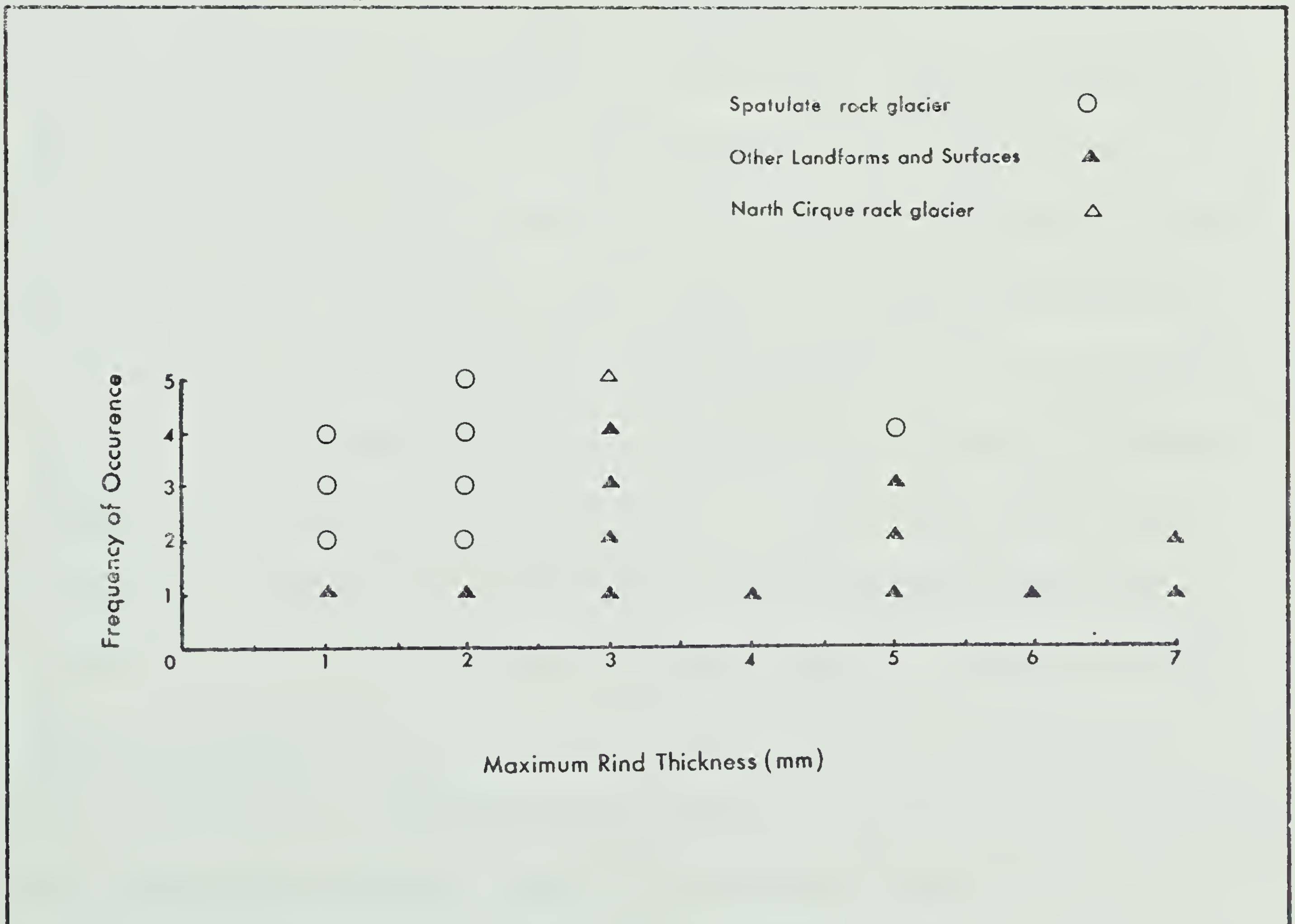


FIGURE 5.4 Histogram of weathering rind thicknesses in the study area.

phase dates from the early 1700's (250 BP) at latest. After the latter phase there is a scarcity of maximum *R. geographicum* thalli suggesting either a period which was not conducive to lichen colonization, or one which was annulled by subsequent geomorphic activity. The oldest phase of colonization has a lichenometric estimate of ~800 BP (1150 A.D.) but, as discussed in Section 5.1, this must be considered as a minimum estimate.

Figure 5.4 is a histogram of the weathering rind thicknesses found in the study area (Tables 4.1, 4.2, 4.3, 4.6). This data is more limited than the lichen data since many of the more recent deposits lack rinds (see Sec. 5.2). However, excluding zero rind values, Figure 5.4 shows a complete range in rind thicknesses from 1 to 7 mm, with slight peaks at 3, 5 and 7 mm. When the additional data from the rock glaciers is included, the large number of 1 and 2 mm rinds on the spatulate rock glacier cause the 3 mm peak to shift down and widen. The result is that there is a distinct modality at 1-3 mm and 5 mm. These characteristic thicknesses may be interpreted as spans of time when exposure, stabilization, and reduced mechanical weathering of surfaces had taken place. Based on the relationship described in Section 5.1, these peaks occur when the lichens are ~60-110 mm and ~160 mm in size, and can in turn be lichenometrically dated at 180 BP (1770 A.D.) to 670 BP,and ~1000 BP, respectively. These dates do not coincide with the dates indicated by the peaks on the lichen size histogram (Fig. 5.3), and may indicate that the relationship between rind thickness and lichen diameter, described in

Section 5.2, is inaccurate.

The above discussions show that there is a certain amount of overlap between the lichen and weathering rind data. In order to obtain a clearer picture of the relative ages of the sample sites, the lichen and weathering rind data were combined. Previously, it was noted that there are direct relationships between lichen size, percentage of surface cover by lichens, and weathering rind thickness, but that these values could vary considerably within any given site. In order to overcome this problem and to obtain a rough estimate of the chronological order of the sites, each of the relative age criteria were divided into 8 classes which were assigned values from 1 to 8. In this way a total value could be obtained for each site which was the sum of all the relative age criteria. The oldest features would have the largest totals while the youngest would have the smallest. Table 5.1 summarizes these findings and lists the sites in order of decreasing age. Although the above method may be a gross simplification when considering all the factors involved, it is noteworthy that the chronology obtained does not conflict with the relative ages of these sites, as indicated by topographic position and cross-cutting relationships. In addition, these data do point out various features which may be of similar age.

5.4 Soil Morphology and Time

The primary purpose of the soil studies, discussed in Chapter III, was to determine whether differences existed between soils found on

Table 5.1 Weathering rind and lichen values (ranked 1 to 8) for selected landforms and surfaces in Whistlers Creek Valley. The features are listed in order of decreasing age suggested by the sum of these values.

	Lichen size	% Total Lichen Cover	Rind Thickness	Sum
Tor (Marmot Mtn.)	8	--	7	15*
Spatulate rock glacier	8	8	6	22
Bedrock surface (between Indian and Whistlers Passes)	7	8	6	21
Lateral moraine 1. (North Cirque)	8	8	4	20
Rock glacier (adjacent to spatulate rock glacier)	7	7	6	20
Scree (east of North Ridge)	4	8	8	20
Block debris (Whistlers Pass)	5	--	8	13*
Lobate rock glacier (North Ridge)	7	7	3	17
Terminal moraine (North Cirque)	4	8	4	16
Terminal moraine (North Cirque)	4	8	4	16
Bedrock surface (in front of South Cirque rock glacier)	4	8	3	15
North Cirque rock glacier (lower lobe)	6	4	4	14
Bedrock surface (adjacent lateral moraine 2)	3	8	2	13
South Cirque rock glacier (lower lobe)	4	8	1	13
Spatulate rock glacier (incipient lobe)	4	6	1	11
Scree (above lateral moraine 2)	2	6	1	9
South Cirque rock glacier (upper lobe)	3	2	1	6

Table 5.1 Continued

	Lichen size	% Total Lichen Cover	Rind Thickness	Sum
Lateral moraine 2. (North Cirque)	3	1	1	5
North Cirque rock glacier (upper lobe)	1	1	1	3
Lateral moraine 3. (North Cirque)	1	1	1	3

*age assumed from topographic position and available data.

surfaces which were thought to be of different relative ages. Since the least differentiation of soils occurs on rock glaciers and protalus deposits (Mahaney, 1974), these surfaces were avoided. Soil pits one and two (S-1, S-2) were located below tree-line in open stands of Alpine fir (Fig. 3.2). S-1 lies on the valley bottom (elev.~2230 m), whereas S-2 is situated on the bedrock ledge (elev. ~ 2250 m) immediately downvalley of the rock glacier exiting from the South Cirque. Soil pit S-3 was located above treeline (elev. ~ 2480 m), on the floor of the North Cirque, in front of the rock glacier (Fig. 3.2). According to their topographic positions, soil pits S-1 to 3, respectively, lie on surfaces of decreasing age, and should, therefore, reflect some changes in time dependent characteristics within the solum. It must be noted, however, that some of the differences in these soils may not necessarily reflect different ages but only variations in the local microenvironment. Each soil pit is at a different elevation, furthermore, S-1 and S-2 are below tree-line while S-3 is above tree-line. There is also evidence of an old forest fire in the study area. The past occurrence of such fires may have affected pedogenesis, as it does have an effect on rock weathering (cf. Blackwelder, 1927). With this in mind then, the description and taxonomy of the soils are given in Table 5.2. S-1 and S-2 are Brunisols while S-3 is a Regosol. Both of these soil types are usually regarded as either juvenile soils, being precursors to others, or they may also be in a steady state where accretion or erosion equals the rate of pedogenesis.

5.4.1 Soil Horizons

Table 5.2 Description of the soil profiles.

(S-1) Orthic Dystric Brunisol

Horizon	Depth (cm)	
LFH	2 - 0	Loose litter of fresh to partially decomposed vegetative parts.
Ah	0 - 6	Very dark reddish brown (5YR 2/3 m), dull brown (7.5 Y/R 5/3d) sandy loam; amorphous; very friable; plentiful, very fine and fine roots; abrupt, wavy boundary; 4 to 8 cm thick.
Bm	6 - 20	Olive brown (2.5Y 4/6 m), light yellow (2.5Y 7/3d) sandy loam; weak, single grain; friable; few, fine roots; clear wavy boundary; 17 to 20 cm thick; pH 4.4.
Cg	20+	Grayish olive (5Y 5/3 m), light gray (5Y 7/2 d) gravelly, sandy loam; moderate, coarse subangular blocky; friable; some angular stones; pH5.

(S-2) Gleyed Dystric Brunisol

Horizon	Depth (cm)	
LFH	1 - 0	Litter of fresh to partially decomposed vegetative parts.
Ah	0 - 3	Dark brown (10 YR 3/3 m), grayish yellow (2.5Y 6/2d) sandy loam; amorphous; very friable; abundant, fine and very fine roots; clear, wavy, boundary; 2-4 cm thick.
Bmgj	3 - 19	Grayish olive (5Y 6/2 m), grayish yellow (2.5Y 7/2d) sandy loam; weak fine granular; plentiful fine roots; clear wavy boundary; 14-18 cm thick; pH 4.1.
Cg	19+	Dull yellowish brown (10 YR 5/4 m), light yellow (2.5Y 7/3d) gravelly, sandy clay loam; moderate, fine subangular blocky; firm; some angular stones; pH 4.3.

Table 5.2 Continued.

(S-3) Orthic Humic Regosol		
Horizon	Depth (cm)	
LFH	1 - 0	Litter of fresh to partially decomposed vegetative parts.
Ah	0 - 11	Olive black (5Y 2/2 m), brown (7.5 YR 4/3d) sandy loam; amorphous; slightly hard; abundant, fine to very fine roots; abruptly wavy boundary; 11 cm thick; pH 4.4.
C	1H	Yellowish brown (10 YR 5/6 m), light yellow (2.5Y 7/4d) very gravelly, sandy clay loam; single grain; loose; angular, gravelly; pH4.

Both S-1 and S-2 have similar horizons. Moist conditions from a high water table or imperfect drainage and indicated by gleying in the lower horizons (Table 5.2). Gleying is more extensive in S-2, even though it is not in a local depression. This may be the result of its proximity to the rock glacier since the large amounts of water emerging from its front may affect the local ground water table (see Sec. 4.5.2). S-1, which by position is considered to be the oldest soil, has thicker horizons and greater depth than S-2. On the other hand, S-3, the youngest soil, has the least development in depth, but the Ah horizon is much thicker compared with S-1 and S-2. This may be the effect of the different microenvironment in which S-3 is found (i.e., higher elevation, different vegetation).

5.4.2 Particle Size

The grain size distributions within the soils are given in Figure 5.5a, and cumulative grain size curves are in Appendix A. Of the <2 mm size fraction, sand (2-.047 mm) dominates the composition of the subsurface horizons of all the soils. The values range from 53 to 63 percent (Fig. 5.5a). Higher percentages of silt (.047-.002 mm) relative to sand and clay in the surface horizons may indicate a certain amount of aeolian deposition. Higher amounts of clay in the subsoils may indicate *in situ* weathering, or translocation by illuviation (Mahaney, 1974). In soils S-1, S-2, and S-3 no definite distinctions of age could be made on the basis of particle size variations. However, the high percentage of sand (52%) in the surface horizon of S-3 compared to S-1 and S-2 (41 and 42% respectively) may indicate a

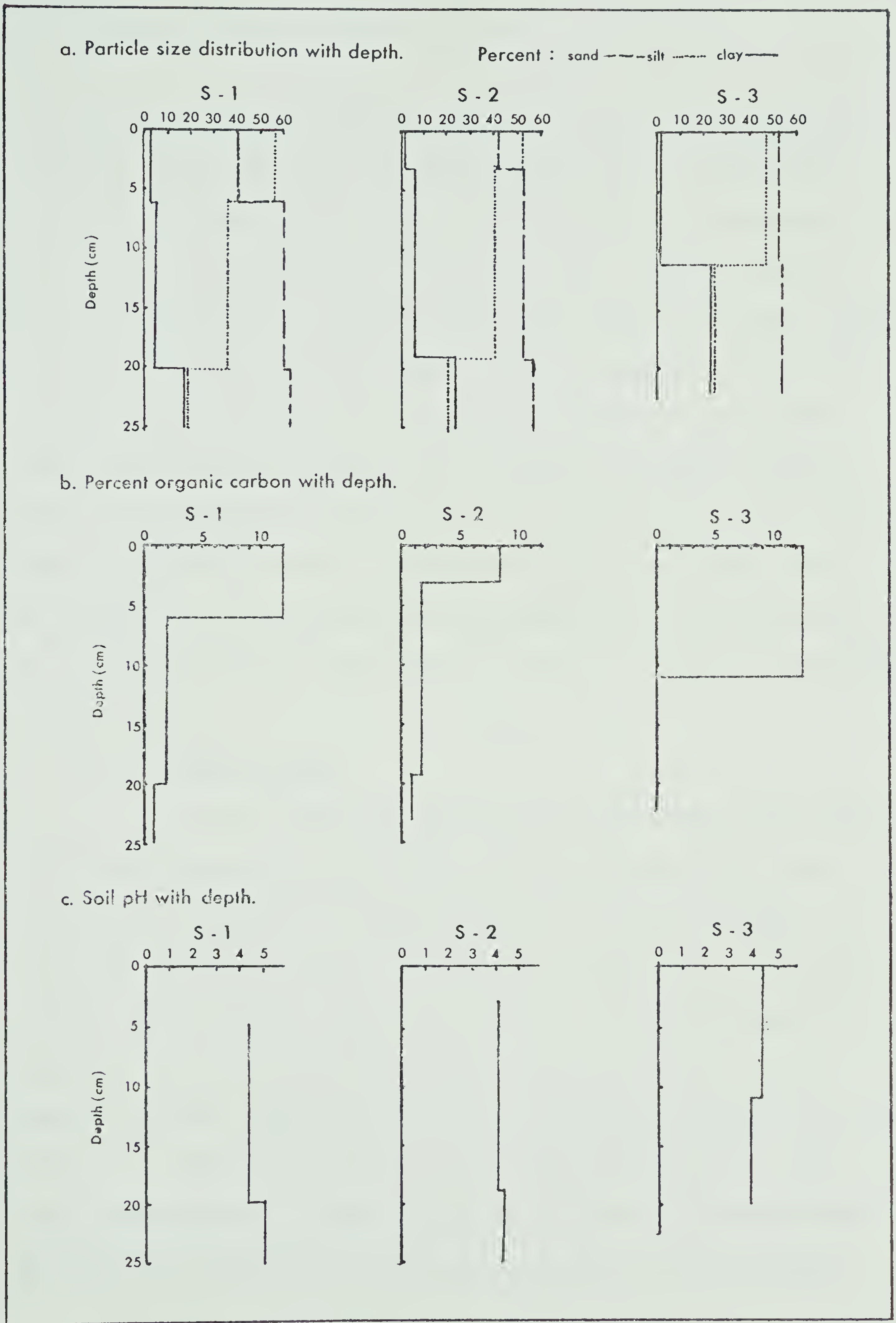


FIGURE 5.5 Some characteristics of the soil profiles.

more youthful soil. In addition, when plotting the texture classifications of the soils (Fig. 5.6), there appears to be a tendency for increased loaminess in the A horizons through time. For comparison, two additional samples were plotted in Fig. 5.6. One was a sample of the fine matrix material exposed on the front slope of the South Cirque rock glacier and the other was taken from the center of a turf-rimmed net (location, Fig. 3.2). The rock glacier material is very sandy (83%) with a low clay content (3%) indicating the youthful characteristics of the unstable slope. The material from the net, on the other hand, is high in clay (41%) and relatively low in sand (25%). This may suggest that the material has been reworked and frost-churned for a long period of time, as the surface on which it is found is considered old (Sec. 5.5).

5.4.3 Organic Carbon

A certain amount of variation in organic carbon (and therefore organic matter) was found within the soils (Fig. 5.5b). When comparing S-1 and S-2, S-1 shows a systematic increase of organic carbon in the Ah horizon over S-2, suggesting an older soil. These soils, however, show a similar dispersal of organic carbon in the subsurface horizons. S-3 does not fall into this sequence as it has a large amount of organic carbon in the A horizon (slightly more than S-1), and there is no evidence of dispersal in the subsurface horizon. Again, these characteristics probably reflect the different microenvironment in which S-3 is found (i.e., above the tree-line), rather than age.

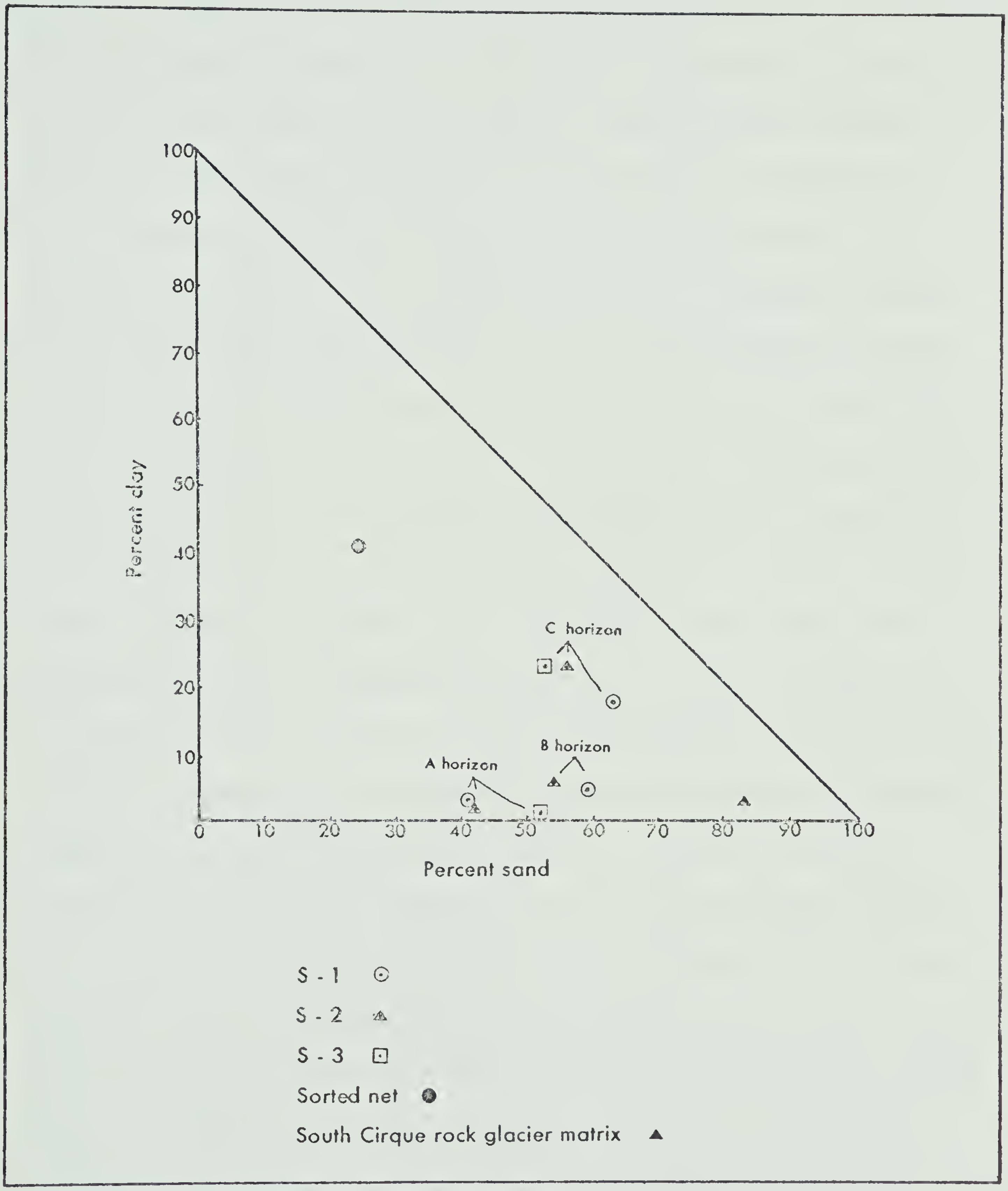


FIGURE 5.6 Textural classification of the soils.

5.4.4 Clay Minerals

The results of the clay mineral analyses for the soil profiles are shown in Table 5.3. Similar clay assemblages for the C horizons in all three soil pits suggest a common parent, geologic material. This parent material is characterized by a dominance of mica, secondary amounts of chlorite, and possible kaolinite. The presence of smectite in S-3 is unusual since it is thought to represent poorly leached, alkaline soils, with a high pH (Mahaney, 1974). It must, therefore, be inherited. According to the previously described relationships (Sec. 3.2.3, E), clay mineral alterations, since deposition of the parent material (C horizons), should be evident in the overlying horizons. In S-3, no clay mineral alterations are indicated, suggesting a youthful soil. S-2, on the other hand, shows weathering of mica or degradation of chlorite. This is indicated by a notable increase in a chloritic intergrade in the B horizon over the C horizon. However, this chloritic intergrade collapses at $300-550^{\circ}C$, suggesting a poorly developed, or juvenile, interlayer material. Nevertheless, the results indicate more advanced weathering than S-3. S-1 also has a chloritic intergrade in the B horizon, but it is less abundant than in the B horizon of S-2. However, this chloritic intergrade persists with heat treatment up to $550^{\circ}C$ which suggests a better formed clay mineral than in S-2. This may, therefore, indicate a slightly longer interval of weathering. In conclusion, the clay mineralogy of the soil horizons suggests progressively more weathered soils from S-3 to S-1 respectively.

Table 5.3 Clay mineral analyses (<2 μ fraction) of the horizons in Table 5.2. Amounts indicated: dominant (XXX); medium (XX); minor (X); trace (tr.); possibly (poss.).

Sample: Site	Sample: horizon	Depth (cm)	Mica	Hydrous Mica	Vermiculite	Chlorite	Chloritic Intergrade	Kaolinite	Smectite	Additional minerals
S-1	Bm	6-20	XXX			XX	X	poss.		
S-1	Cg	20+	XXX	tr.		XX		poss.		Feldspar (X), Calcite (tr.)
S-2	Bmgj	3-19	XXX	X		X	XX	poss.		
S-2	Cg	19+	XXX		tr.	XX	tr.	poss.		Feldspar (X)
S-3	Ah	0-11	XXX			XX	tr.			
S-3	C	11+	XXX			XX	tr.	poss.	tr.	Feldspar (X)

5.4.5 Soil pH

The soil reaction for the samples is very acidic ranging from pH 4 to 5 (Fig. 5.5c), hence S-1 and S-2 are classed as Dystric Brunisols. The data indicate slight increases in pH with depth and a slight decrease with depth in S-3. No conclusive interpretations of soil age could be made from this data.

5.4.6 Conclusions

The above data show a number of differences within the soil profiles studied. These variations may reflect different ages of the soils, however, they could also be caused by differing microenvironmental conditions. In addition, it is not known if the differences in certain characteristics, such as particle size and organic carbon, are not merely due to random variations. Nevertheless, it appears some characteristics, such as the extent and depth of soil horizon development, in addition with the alteration and development of clay minerals, may be useful for relative age differentiation. However, only a greater number of soil pits could prove the validity of these results.

5.5 Glacial and Periglacial History

A provisional chronology of Holocene glacial and periglacial events in the Whistlers Creek Valley can be outlined on the basis of the evidence presented in the previous sections. The Holocene begins in this area with the retreat of the Cordilleran Glacier Complex at the end of the late Wisconsin glaciation. Radiocarbon evidence suggests that the continental divide (~15 km southwest of the study area) was ice free ~10,000 BP (Luckman and Kearney, 1978). At some unknown time in

the past, Cordilleran distributory ice from the Miette Valley entered upper Whistlers Creek and deposited the large erratics and a thin mantle of till (Event I). As the main ice downwasted and retreated to the northwest, ice tongues occupied the valley, causing ice marginal channels to form along the valley walls. These channels, coupled with the structure of the bedrock, formed large bedrock terraces on the flank of Indian Ridge. Accelerated subaerial weathering of the newly exposed surfaces, and extensive mass wasting of the debris left by this ice, led to the formation of tors and blockfields. Tors, however, may be found on surfaces of great antiquity (Dyke, 1976, 1978). The surfaces at, and above, the location of these tors then (Fig. 4.1), may not have been inundated during the last major glacial. The two moraine-like mounds, one at the mouth of the North and South Cirques, and the other before the crest of Whistlers Pass (Fig. 4.1), may mark recessional stages in the retreat of the main Cordilleran ice from the valley. The subdued rock glacier-like topography in the upper reaches of the study area (Fig. 4.1) may have formed if a remnant ice mass became covered by detritus and subsequently evolved into a rock glacier.

At an unknown time, subsequent to the recession of Cordilleran ice from the valley, soil S-1 began to form. After this indeterminate period of ameliorated climate, relative age criteria indicate a time of intensified periglacial activity which led to the formation of the spatulate rock glacier and numerous talus deposits (Event II). The minimum lichenometric date for this event is > 950 BP, however, it is likely much older given the mature state of the spatulate

rock glacier. During this period, or possibly during a subsequent deterioration of climate, there was an advance of the ice glaciers occupying the North and South Cirques. The maximum extent of this advance is not known, but the lateral moraine in the North Cirque suggests that the two glaciers coalesced as they moved downvalley. Consequently, it may be that the moraine-like feature below the cirques on the main valley floor (Fig. 4.1) is the terminal moraine from this advance, and not a remnant from the retreat of the main Cordilleran ice. If the spatulate rock glacier predates this period, it must have experienced renewed activity at this time since the rock glacier immediately adjacent to it,and the lobate rock glaciers on the North Ridge,were forming. There are two minimum dates on the time of this advance, one is >850 BP based on a maximum lichen thallus on a lateral moraine, and the other is > 350 BP, based on the age of a spruce tree (T-14, Fig. 3.2). With the retreat of the cirque glaciers, the surface where soil S-2 is found was exposed. Two recessional moraines in front of the North cirque rock glacier (Fig. 4.1) indicate at least two stillstands or minor readvances during the recession of the cirque glacier. With evacuation of the cirque floor by the ice, the surface on which soil S-3 is found was exposed.

Evidence for the next period of glacial activity (Event III) occurred with the advance and subsequent stagnation of the ice glacier within the North Cirque, which formed the lower lobe of its rock glacier (>575 BP). There is also evidence of increased production of talus and consequent exposure of fresh surfaces at this time. During this period,

the glacier in the South Cirque may also have advanced and stagnated, producing the lower lobe of the South Cirque rock glacier. However, if this is the case, the oldest part of the lobe must have been overridden by the subsequent advance, since the lichenometric date for the lower lobe is ~1770 A.D. (190 BP), compared to 575 BP for the lower lobe of the North Cirque rock glacier. In any case, the glaciers in the North and South Cirques did not coalesce, as indicated by the preservation of the older, North Cirque lateral moraine (Event II). The advance in the South Cirque, however, truncated the outer section of this lateral moraine, where it exits from the North Cirque. Limited rejuvenation of the spatulate rock glacier, indicated by the formation of an incipient lobe near its head, presumably occurred during this period (lichen dated ~1700 A.D., 250 BP). However, this latter rejuvenation was limited by either an insufficient supply of debris, and/or an insufficiently rigorous climate.

In the last phase of this period (Event III) the upper lobe of the South Cirque rock glacier formed with the last advance and subsequent stagnation of the ice glacier (~1885 AD) on the top surface of the pre-existing rock glacier (~1770 A.D.). At approximately the same time, the ice glacier in the North Cirque also readvanced and overrode the rock glacier below it. Subsequent retreat left a lateral moraine against the north wall of the cirque (lichen dated ~1880 A.D.). Continued downwasting of the retreating ice glacier led to the isolation, and burial by debris,of a remnant ice tongue which subsequently took on the characteristics of the upper lobe of the rock glacier.

Lichen evidence indicates that the surface of this lobe became stable at ~1925 A.D. A smaller lateral moraine, below, and parallel to, the above 1880 A.D. lateral moraine suggests a minor stillstand in the retreat of the ice glacier in very recent times. Since there are no lichens on this moraine,it may be post 1950 A.D. in age. Evidence from air photographs,discussed in Section 4.2.2, also supports this.

In the present environment,the two cirque glaciers are retreating while the downvalley rock glaciers in the North and South Cirques are advancing. On the other hand, the spatulate and lobate rock glaciers are inactive (Fig. 4.1). Evidence from semipermanent snowbanks,as shown on air photographs (Sec. 4.2.3),indicates deteriorating conditions since at least 1950 A.D. Solifluction is very active in the present environment, but must have had periods of reduced activity in the past as have the other glacial and periglacial processes. A number of solifluction features show evidence of reactivation which may be due to the climatic change suggested by the increased persistance of the snowbanks mentioned above.

In conclusion, the above is a tentative outline of the glacial events in Whistlers Creek Valley as indicated by the relative age criteria. These criteria are by no means certain, especially for determining the older events. Therefore, this glacial history will undoubtedly be open for modification and refinement with further work in the area. In addition,the above discussion suggests that the glaciers in the North and South Cirques do not necessarily respond synchronously to climatic changes,despite the fact that they

are very close to one another. This factor must be considered when attempting to correlate events with other areas. Table 5.4 summarizes the glacial and periglacial history for the study area.

5.6 Correlation

The Holocene history of Whistlers Creek Valley was preceded by the recession of a late Pleistocene ice mass of unknown age. The time of the last major Cordilleran ice advance in the Athabasca River Valley, the Obed (Roed, 1975), is not known, but it was tentatively correlated with the Late Portage Mountain advance of the Peace River area which culminated at 11600 ± 1000 BP (Rutter, 1976). The North Saskatchewan River Valley, to the southeast of Whistlers Creek, may have become ice free by 9330 ± 170 BP at Saskatchewan Crossing (Westgate and Dremanis, 1967). The Tonquin Valley, at the continental divide, was ice free by at least 9660 ± 280 BP (Luckman, 1978), based on a basal peat date. This evidence suggests that the general region surrounding the study area was ice free by ~10,000 BP.

Subsequently, an early period of rock glacier development was recognized in the Banff and Yoho National Parks area where Mazama ash (6600 BP) was found in soils forming on the rock glaciers (Osborn and Duford, 1976). Luckman and Crockett (1976) also suggest an early period of rock glacier activity between 6000 and 9000 BP, and they recognize at least two additional periods of rock glacier development prior to the Little Ice Age (1450-1880 A.D.) in the Jasper area. In Whistlers Creek Valley the spatulate rock glacier, and possibly the

Table 5.4 Summary of the glacial and periglacial history of Whistlers Creek Valley.

	Age (BP)	
Event I	>10,000	Cordilleran Ice occupies valley.
		Retreat and stagnation of Cordilleran Ice. Possible formation of Soil (S-1) begins.
	>950	Development of the Spatulate rock glacier.
Event II	>850	Coalescent advance of the North and South Cirque rock glaciers.
	>760	Development of the lobate rock glaciers.
		Retreat of the cirque glaciers (with readvances or stillstands).
		Possible formation of Soil (S-2) begins.
		Possible formation of Soil (S-3) begins.
		Advance of the North Cirque glacier.
	>250	Possible advance of the South Cirque glacier.
		Possible development of incipient lobe on the spatulate rock glacier.
	>580	North Cirque glacier stagnates, North Cirque rock glacier lower lobe developes.
Event III		
	>180	South Cirque glacier stagnates, South Cirque rock glacier lower lobe developes.
	>70	Readvance of the South Cirque glacier.
		Readvance of the North Cirque glacier.
	>65	Stagnation of the glacier and development of the South Cirque rock glacier upper lobe.
	>25 (1925 A.D.)	Stagnation of the glacier and development of the North Cirque rock glacier upper lobe.

lobate rock glaciers against the North Ridge, may represent this early period of rock glacier activity (Event II). A subsequent or contemporaneous phase of activity is indicated by the advance of ice from the North and South Cirques. The presence of the pre-Little Ice Age moraine in the North Cirque, associated with this advance, is rare. Luckman and Crockett (1978) note that in the Jasper area the majority of pre-Little Ice Age, Holocene glacial periods are represented by rock glaciers rather than moraines. The exceptions are the old moraines at Mount Edith Cavell and the Athabasca Glacier. Lichenometric dates attained for the moraines at Mount Edith Cavell range from 724 BP to 1760 BP (Luckman, 1977) and may, therefore, be correlative with Event II (minimum date 850 BP). In addition, Osborn (1975) has lichenometrically dated a late Holocene advance of the Victoria Glacier, Lake Louise,from at least 650 BP.

The formation of the rock glaciers in the North and South Cirques, and the partial rejuvenation of the spatulate rock glacier (Event III), covers a time span from before the Little Ice Age to its culmination. The lower lobe of the North Cirque rock glacier is pre-Little Ice Age (575 BP). The incipient lobe in the spatulate rock glacier (1700 A.D., 250 BP) and the lower lobe of the South Cirque rock glacier (1770 A.D., 180 BP) make up the earliest evidence of the Little Ice Age in the study area. Furthermore, the culmination of the Little Ice Age in the study area is represented by the upper lobes of the North and South Cirque rock glaciers (dated 1925 and 1885 A.D., respectively). The Little Ice Age events at Whistlers Creek seem to correlate well with

events in other areas. The only radiocarbon evidence from the earliest phase of the Little Ice Age is 450 ± 50 BP from the Mount Robson area (Heusser, 1956). The earliest Little Ice Age moraines at Mount Edith Cavell and Penstock Creek were lichenometrically dated at 1705 ± 5 A.D. (245 BP) and 1765 ± 5 (185 BP), respectively, whereas the youngest moraines were dated at 1888 ± 7 A.D. and 1907 ± 5 A.D., respectively (Luckman, 1977). Similarly, Osborn (1975) lichenometrically dated a moraine of the Victoria Glacier at 180 BP. In addition, he dated a recessional moraine from 1897 A.D., based on photographic evidence.

For further comparison, Table 5.4, the glacial history of Whistlers Creek, may be compared with Table 1.1, a summary of Rocky Mountain Holocene events.

CHAPTER SIX

CONCLUSION

The study of the Holocene glacial and periglacial features found within Whistlers Creek Valley shows a diverse set of geomorphic processes at work, with several phases of past activity. This valley displays many of the classical characteristics resulting from alpine glaciation. Two well developed cirques (the North and South Cirque) occur in the upper reaches of the valley and both are presently occupied by small cirque glaciers. The South Cirque also has a small niche glacier located in a crevice in its upper headwall.

Many of the landforms which have resulted from past environmental conditions were tentatively dated using several relative age dating techniques. Cross-cutting relationships between features established an initial chronology of events. Lichenometry proved to be very useful for providing minimum dates on the more recent rock glacier activity (i.e., 250 BP). On older surfaces, isolated *R. geographicum* thalli became exceedingly hard to distinguish, making measurement impractical. Problems of species succession and senescence of the lichen also became evident when attempting to date older surfaces. Nonetheless, gross distinctions of age were made, based on maximum lichen size and their total surface coverage. Relative age determinations, based on differential rock weathering, were limited owing to the resistant nature of the quartzite rocks in the study area. Oxidation rinds were the only weathering feature that could be systematically measured on the rocks, and gross distinctions of weathering zones could also be made. Soil

profiles, located on surfaces of varying ages, showed minor variations which may, however, also be related to varying microenvironments. Nonetheless, slight differences in depth of profile development, horizon thickness, and clay minerals may be related to age. In summary, cross-cutting relationships were the most essential relative age determinants. However, an important point about the lichen and rock weathering studies is that while they were undertaken independently of each other, collectively, the results are in agreement.

The earliest evidence of ice in Whistlers Creek Valley is from a number of erratics and a thin discontinuous drift sheet left by the Cordilleran Glacier Complex, probably of late-Pleistocene age. According to the relative age criteria, the earliest Holocene glacial activity in the study area was manifest in the development of spatulate and lobate rock glaciers. In the present environment, these rock glaciers have stable, gently sloping fronts and are inactive. They issue from talus slopes at the base of steep cliffs and are thought to have developed and flowed via interstitial ice forming within the talus. Complex forms of surface microrelief on these rock glaciers, such as ridges and furrows, are thought to result from differential flow conditions within the ice, and from differential melting of ice rich areas, enhanced by incising meltwater channels. A subsequent phase of glacier activity is recognized by a lateral moraine in the North Cirque. During this period the glaciers in the North and South Cirques are thought to have coalesced and flowed to the valley floor.

The most recent glacial activity in Whistlers Creek Valley is related to the formation of two tongue-shaped rock glaciers immediately

downvalley from the ice glaciers in the North and South Cirques. The development of these rock glaciers is linked with successive advances and stagnations of the adjacent ice glaciers, coupled with the supply of abundant debris by rockfalls and avalanches from surrounding cliffs. In both cases bedrock obstructions were instrumental in cutting off the supply of ice to the glacier fronts which were, in turn, quickly buried during stagnation. The insulative properties of the debris mantle, together with its mass, allowed these buried ice cores to advance while the upvalley ice glaciers entered a period of recession. Steep and unstable fronts, lying at the angle of repose, indicate that these rock glaciers are active. In addition, the South Cirque rock glacier is advancing onto a vegetated surface tilting trees in the process. The minimum rate of advance that has occurred in this latter area is 2.2 $cmyr^{-1}$ for the last 50 years. Comparison of the obtained lichen data with a lichen growth curve (Luckman, 1977) indicates that the earliest period of development occurred ~1770 A.D. (180 BP) for the South Cirque rock glacier, and at least 575 BP for the North Cirque rock glacier. Both of these rock glaciers have younger lobes overriding the earlier lobes indicating a period of rejuvenation. The youngest phase of rock glacier activity occurred at ~1885 A.D. in the South Cirque and ~1925 A.D. in the North Cirque. The recent activity within these Cirques correlates well with Little Ice Age glacial advances in nearby areas (Luckman, 1977; Osborn, 1975).

Solifluction is a widespread process operating in the study area, producing numerous landforms such as stone and turf banked lobes and

terraces. Patterned ground is found in the form of stripes, polygons, steps, nets and hummocks. Large ploughing blocks up to 4 m in length have travelled on low angle, vegetated slopes, leaving furrows of up to 20 m in length. A noteworthy observation is that the present alpine environment favors the periglacial processes which produced these landforms in the past. In fact, some evidence suggests rejuvenation of solifluction forms. Limited information from air photographs suggests that permanent snowcover has increased since at least 1950 A.D.

In conclusion, the study presents an initial effort in describing the geomorphic activity of Whistlers Creek Valley through time,and a number of suggestions for further study can be made. Firstly, the study area contains a number of rock glaciers of apparently different genetic types and degrees of activity. Since these rock glaciers are relatively close to one another, this area would be ideal for more detailed research to determine their mode of flow and factors of formation. Some of these studies could include long term measurement of: horizontal and vertical movements, meltwater discharge, and debris accumulation which could be compared with mass balance studies on the nearby glaciers. Seismic data might indicate the presence and extent of ice within the rock glaciers. A second area of research could focus on refining the chronology within the valley. This could be done if strategically relevant organic deposits or ash layers were found. A sediment core from the bottom of the lake,in the lower portion of the North Cirque valley,might also reveal past environmental fluctuations from its pollen and sediment characteristics. This lake

basin might also provide useful radiocarbon dates on the approximate age of local deglaciation. Thirdly, the large size of the ploughing blocks warrants further research. Movement studies could be made by setting up control points from which measurements could be made from year to year or season to season. In addition, excavation of the trailing furrows may reveal a record of buried organic material which would be suitable for radiocarbon dating. Similar observations could also be made in relation to some of the large solifluction lobes. Lastly, the relationship of the large glacial erratics deposited by Cordilleran ice (from the main Miette Valley) could be further studied in terms of their time of deposition and their implications for maximum vertical extent of ice.

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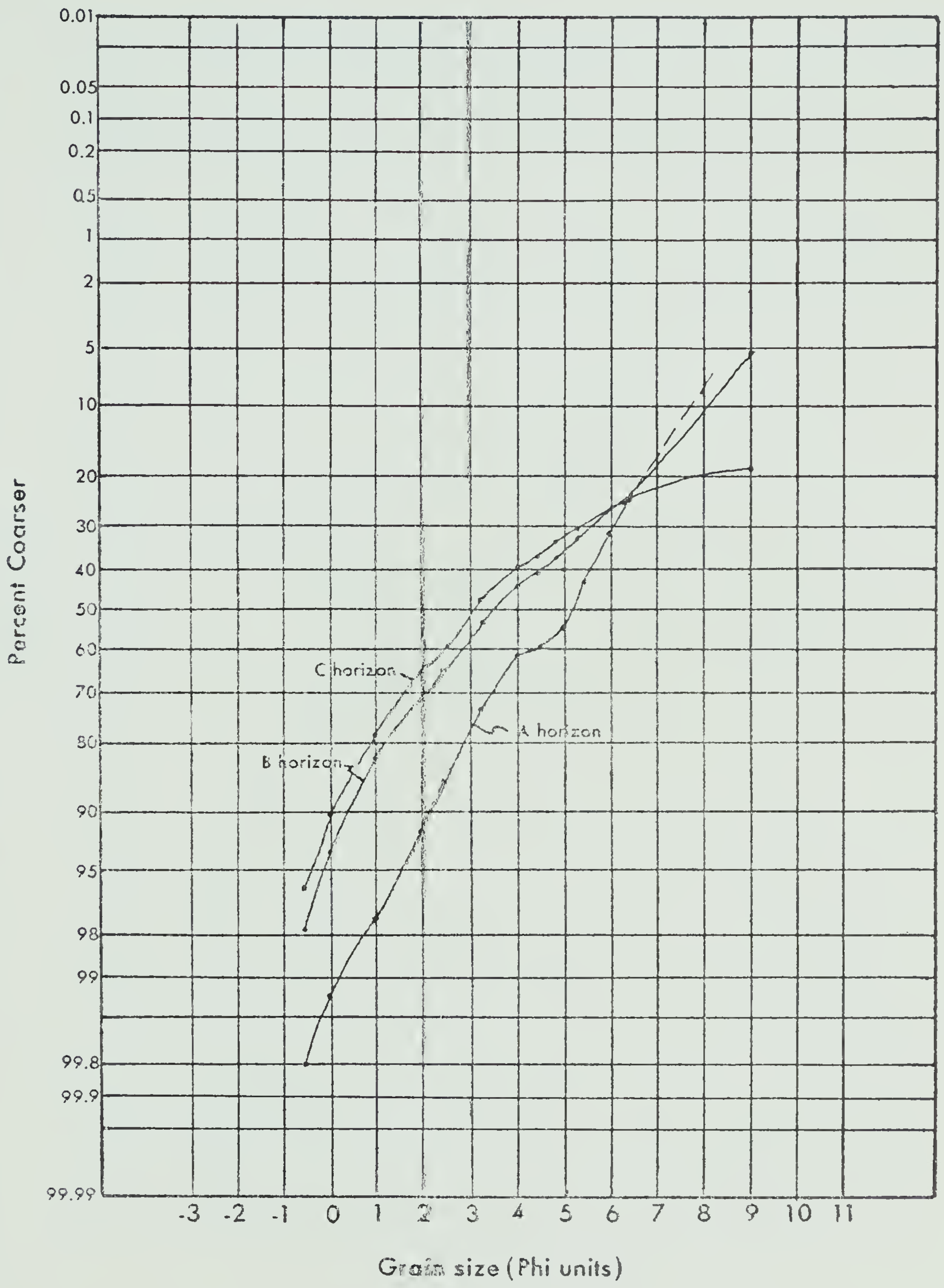
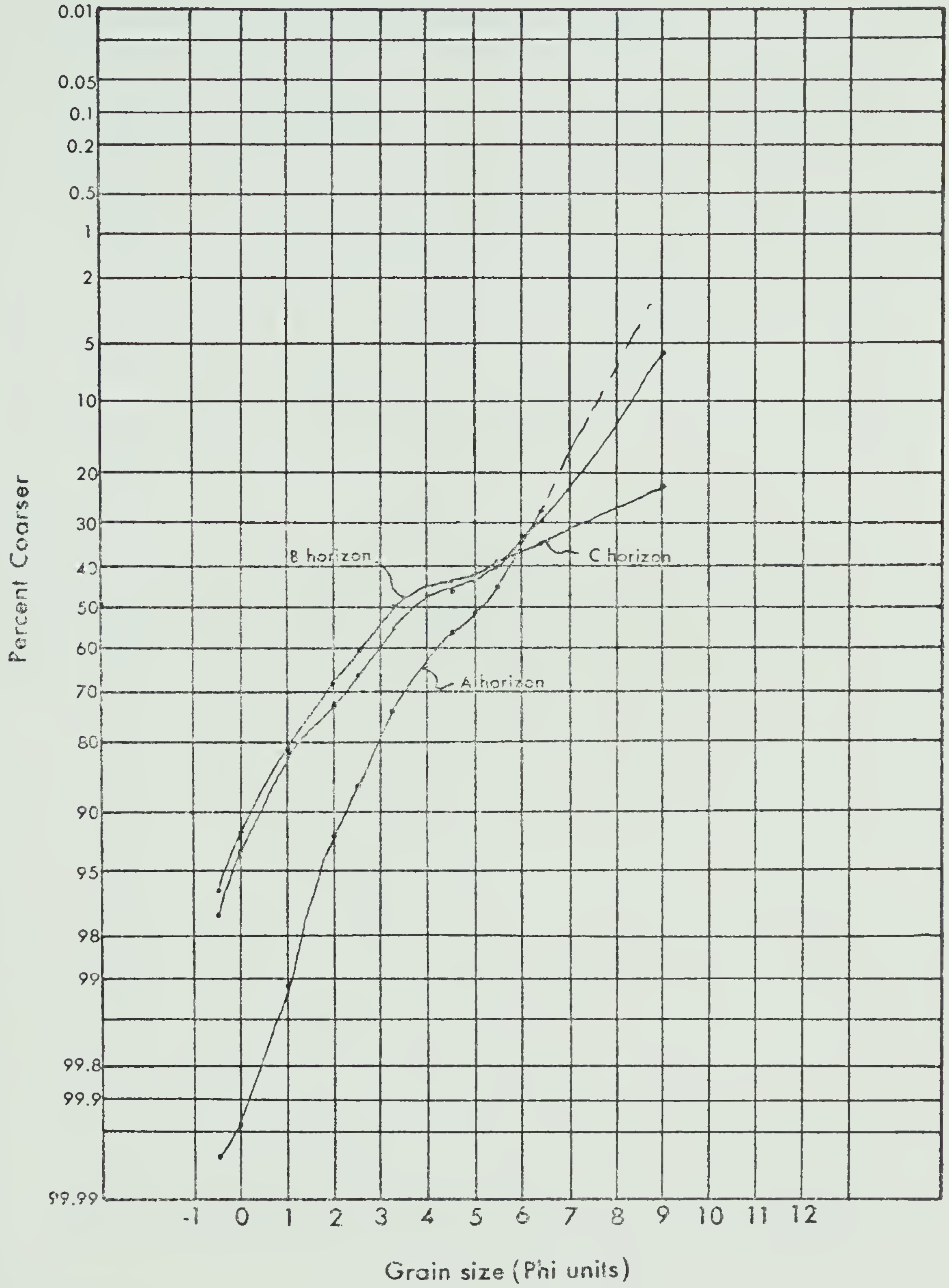


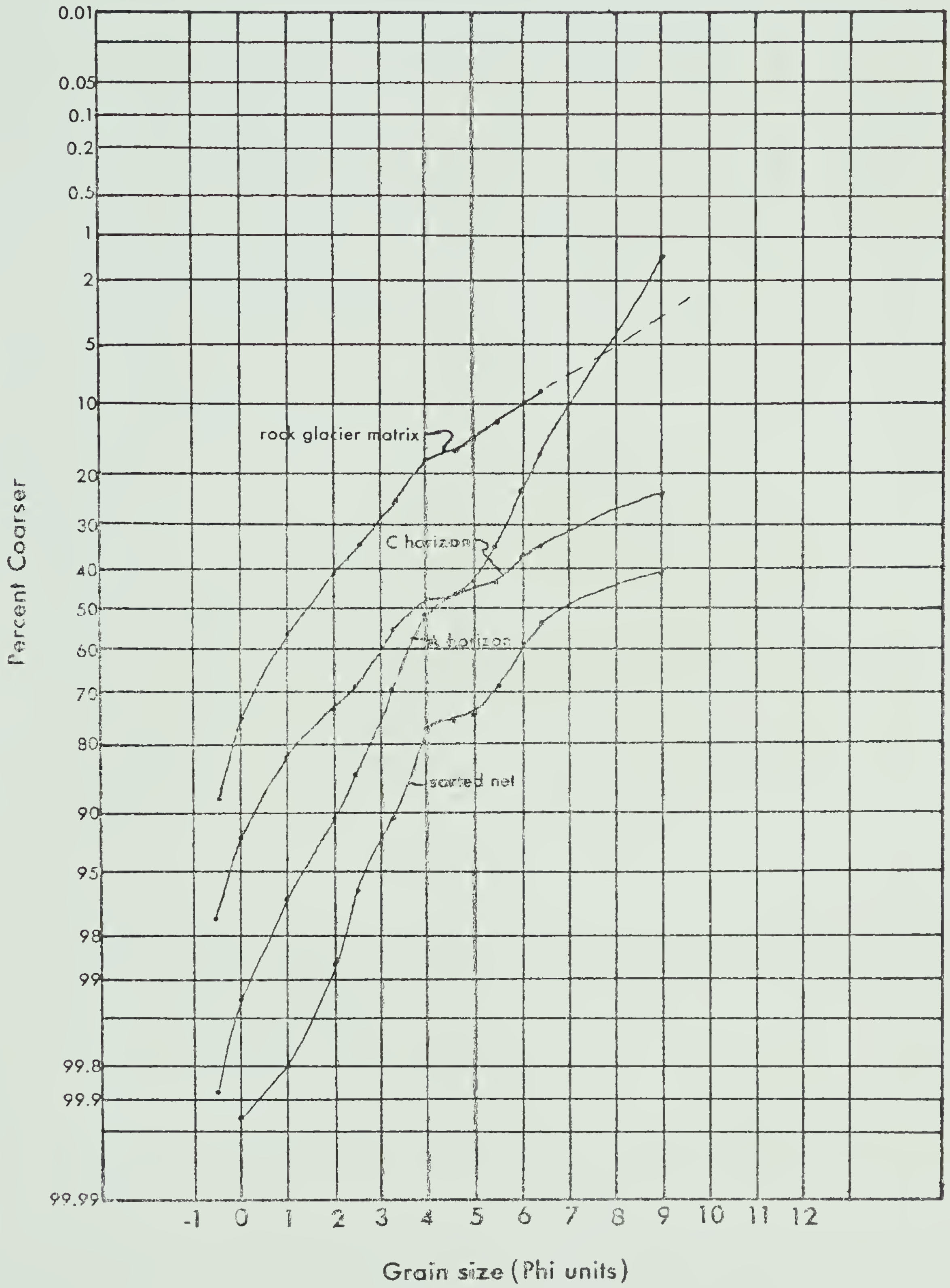
Particle size analysis for soil profile S - 1.

APPENDIX A



Particle size analysis for soil profile S - 2.

APPENDIX A



Particle size analysis for soil profile S - 3, the South Cirque rock glacier matrix , and a sorted net.

APPENDIX A